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REFERENCE

Lane, R. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

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MAY 1958

ISOSTASY AND A THEORY FOR THE ORIGIN OF GEOSYNCLINES*

K. JINGHWA HSU

ABSTRACT. The crust of the earth is in a state of approximately isostatic equilibrium and its structure can best be described by an Airy-Heiskanen model, with regions of thin crust as earth depressions.

The isostatic equilibrium of the crust can be locally disturbed by folding, thrusting, and/or other tectonic disturbances. This paper presents a theory that geosynclines are earth depressions created after tectonic disturbances when the isostatic equilibrium is restored by crustal faulting. Geosynclines are simply regions of thinner crust. Further geosynclinal subsidence is induced by sedimentary load and by compaction of sediments.

The thickness and the nature of the geosynclinal sediments depend primarily upon the thickness of crustal columns underlying a geosyncline. Thick deep-water sediments are deposited in basins on outer continental margins, where the crust is thin. Thin shallow-water sediments accumulate in intracratonic basins.

A normal geosynclinal cycle starts with deep marine sediments followed by shallow marine and deltaic deposits, and ends with "blanket-type" shelf clastics and carbonates. In places, however, the rapid rate of alluvial sedimentation prevents the development of a normal geosynclinal cycle, and regions of thin crust are overlain with thick continental deposits (e.g., Triassic Newark series).

The sedimentary and tectonic histories of southern California and of the central Appalachians are briefly reviewed. The initiation of geosynclinal cycles can be related to periods of local tectonic disturbances.

INTRODUCTION

The origin of geosynclines[1] is a matter of much speculation. The cause of geosynclinal subsidence has been attributed to compression due to contraction of the earth (Dana, 1873), to tension because of the expansion of the earth (Bucher, 1933), or to crustal downwarp by subcrustal convection currents (Vening Meinesz, 1954; Griggs, 1939; Hess, 1939). Sedimentary load has been considered responsible for geosynclinal subsidence (Fisk, 1952; Russell, 1940). Others related the origin of geosynclines to compression induced by isobaric movements of the crust (Gunn, 1947; Officer et al., 1957).

Except for regions that have undergone recent tectonic disturbances, the earth's crust is in a state of approximately isostatic equilibrium (Daly, 1941). That the earth's crust will seek to re-establish isostatic balance after the equilibrium has been disturbed is also generally recognized. Crustal blocks of different elevations above or below sea level will form when isostatic equilibrium is re-established if the crustal thickness differs from place to place. Thinner crust underlies surface depressions. The writer will demonstrate by deductive reasoning that these depressions, initiated by isostatic readjustment of crustal blocks, should develop into different kinds of deposi-

* Publication No. 144, Shell Development Company, Exploration and Production Research Division, Houston, Texas.

[1] Geosynclines as used in this paper refer to depressions of the earth's crust in which sediments of excessive thickness have been deposited (Dana, 1873; Bucher, 1933; Kay, 1951).

tional basins known to us as geosynclines and intracratonic basins. It will be further demonstrated that the sedimentary history of some of the better known geosynclines is consistent with the theory.

The conclusions of this paper are deduced from the two simple premises that parts of the earth's crust are thinner than others and that the earth's crust has always sought to establish or re-establish isostatic equilibrium. The author has refrained, however, from speculating more on the origin of geosynclines than by emphasizing that they are simply regions of thinner crust, as he did not wish to deduce from speculative assumptions that cannot be proven. The reason why the crust is "thinner" or becomes "thinner" in certain regions so as to develop into a geosyncline is a matter of speculation. Certain geosynclines might have been developed on part of the oceanic crust, which had always been thin but only then started to receive considerable supply of sediments. The crust under other regions might be thinned by tectonic processes, such as subcrustal erosion of the crust (Gilluly, 1955), or removal of crustal materials from a "blister" (Rich, 1951), or removal of crustal material by surficial tectonic transport. Whether there can be a unique solution for the fundamental cause of crustal thinning must await future studies.

STRUCTURE OF THE EARTH'S CRUST AND ISOSTASY

Geodetic studies indicate that the earth's crust[2] is for the most part in a state of approximately isostatic equilibrium. Exceptions are in regions of recent tectonic deformation (Daly, 1941). Seismic evidence gives some support to Airy's *root hypothesis* of isostatic compensation. Crustal thickness is about 5 km under the ocean, 35 km under the Precambrian shield areas, and 50 km under the Sierra Nevada (Press, 1956).

A schematic model of a segment of the crust through North America, constructed on the basis of Airy's *root hypothesis*, is shown in figure 1a. A "standard crust" is assumed to have an average density of 2.84, and the mantle below the Mohorovičić discontinuity a density of 3.27 (Worzel and Schubert, 1955).

Airy's *root hypothesis* is an over-simplified concept. The crust under seas and oceans is in fact covered by sediments of varying thickness as shown by seismic studies (Ewing and Press, 1955). If a density of 1.03 is assumed for sea water and 2.3 for the veneer of unconsolidated sediments, a slightly modified crustal model is obtained (figure 1b modified after Worzel and Schubert, 1955). On land, crustal blocks with excess mass above sea level are assumed to be compensated by "roots." The crustal model (fig. 1b) has been constructed largely on the basis of seismic information on the crust of eastern North America and underlying the oceans (Ewing et al., 1954; Raitt, 1956; Tatel et al., 1953; Hodgson, 1953). However, crustal thicknesses of the western part of the continent and the continental margin (except under the Sierra Nevada) as calculated from seismic information are thinner than those predicted by the model (Tatel and Tuve, 1955; Press, 1956; Gutenberg, 1957).

[2] Crust as used in this paper refers to the silicic layer bounded below by the Mohorovicic discontinuity.

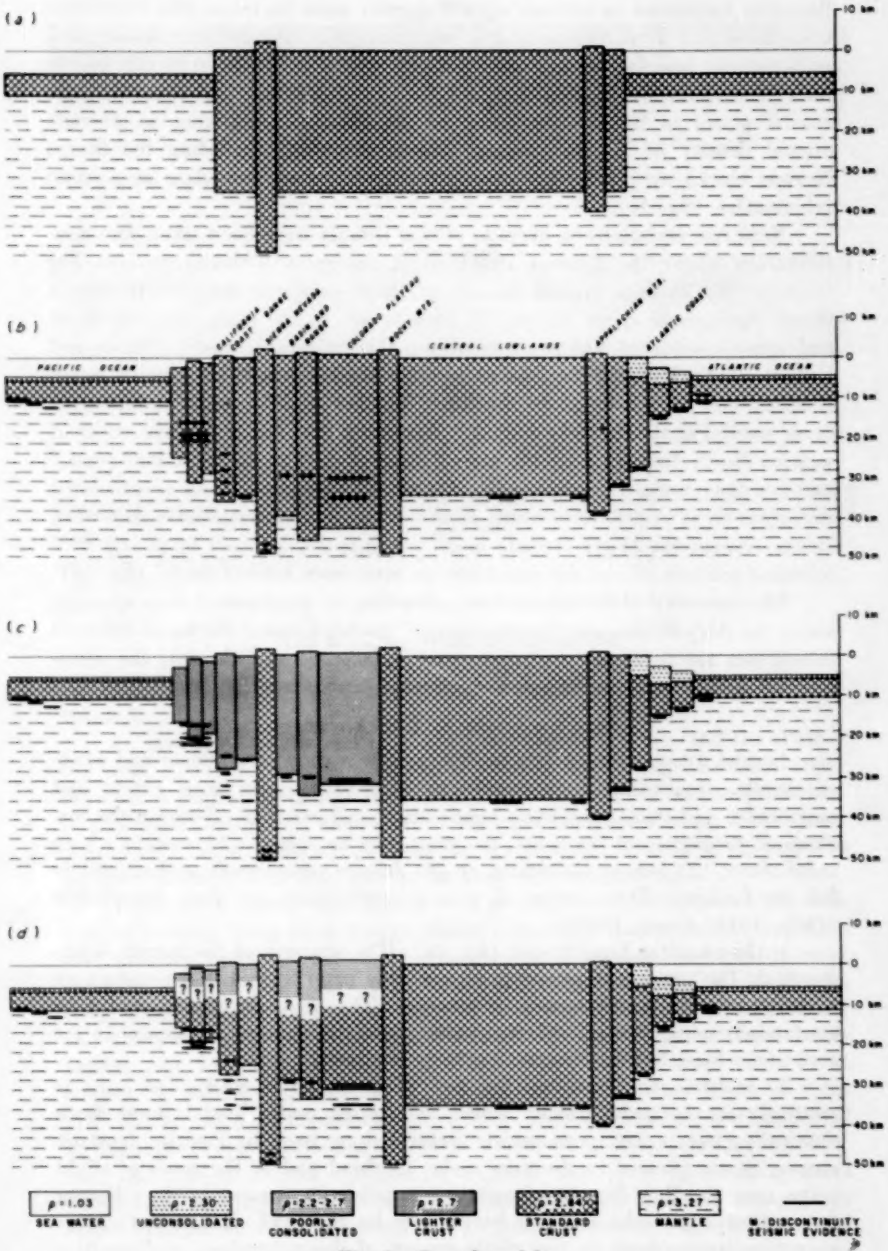

Fig. 1. Crustal models.

Regional variations in average crustal density must be taken into consideration. Tatel and Tuve (1955, p. 50) suggested that "the average density is 7 or 8 percent less for the entire column of crustal rocks down to the mantle under the Colorado Plateau than for the crustal column under the Atlantic coastal plain, or the Appalachians, or the Mesabi." Assuming an average crustal density of 2.7 for western North America (excluding the Sierra Nevada block) we can further modify the model to accommodate the seismic information (fig. 1c).

Regional variations in average crustal density have been taken into consideration where the layered structure of the crust is better known. For example, the average crustal density of the continental margin off eastern North America is about 2.6 to 2.7. This is due to the thick unconsolidated and semiconsolidated sediments overlying the "standard crust" (Officer and Ewing, 1954). Perhaps the smaller average crustal density of western North America can be attributed to the thick Tertiary and Mesozoic sedimentary cover overlying the "standard crust." Assuming that the crystalline rocks of western North America have the same density of 2.84 as the "standard crust," and that these rocks are overlain by varying thicknesses of poorly consolidated sediments so as to give an average density of 2.7 for western North American blocks (except the Sierra Nevada block which is not overlain by poorly consolidated sediments), we can construct an even more refined model (fig. 1d).

The structure of the crust, then, according to geophysical data approximates the Airy-Heiskanen "mountain-root" model. Crustal blocks of different thicknesses are in a state of approximately isostatic equilibrium; the elevations of the blocks above or below sea level depend primarily on the thickness.

CRUSTAL EQUILIBRIUM AND GEOSYNCLINES

Crustal Equilibrium.—The earth's crust is resting on a denser and weaker mantle. Depending upon conditions of loading (i.e., areal extent and magnitude, and time) and upon temperature and pressure effects on the mechanical properties of the rocks, the crust may be able to (1) sustain a load indefinitely, (2) behave elastically, or (3) deform permanently and ultimately fail by faulting. Three types of crustal equilibrium are thus conceivable (Daly, 1941; Gunn, 1949):

1. Mechanical Equilibrium (fig. 2a): The strength of the mantle is immaterial. The crust is supposed to be perfectly rigid so that it can sustain an added load with no distortion whatsoever. The load, therefore, cannot be isostatically compensated; a large positive isostatic anomaly is present in the area under load, but the Bouguer anomaly is absent.

2. Isobaric Equilibrium (fig. 2b): The mantle is weak. The crust is relatively strong and perfectly elastic. The crust under load is bent down elastically, but is not permanently deformed or broken. Positive isostatic anomalies are present in the area under the load and in the areas of slight bulge (see fig. 2b). Negative isostatic anomalies are present in the adjacent depressions where the crust is bent down by the load. Negative Bouguer anomalies are present in the whole region of down bending, and positive

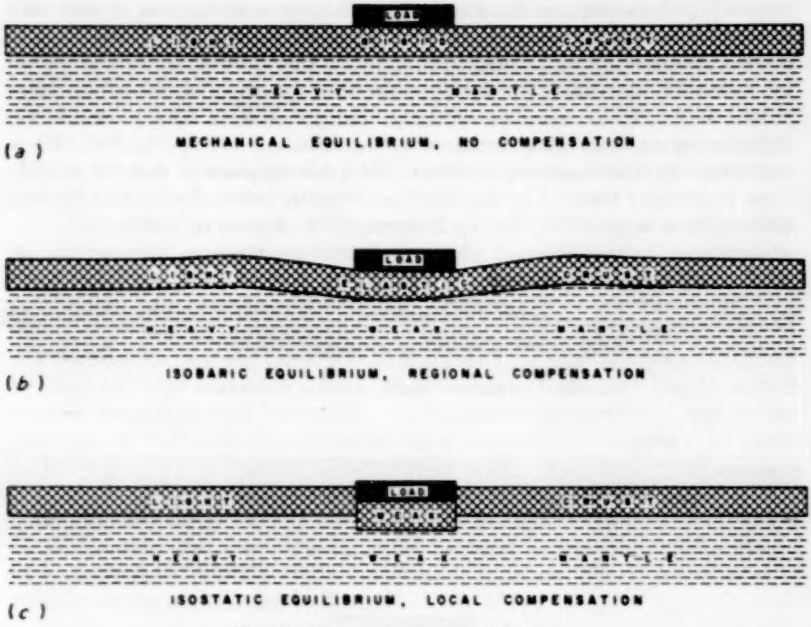

Fig. 2. Types of crustal equilibrium.

Bouguer anomalies are found in the bulged areas where the crust is deflected upward.

3. Isostatic Equilibrium (fig. 2c): The mantle is weak, and the crust is too weak to support the load without flowing or fracturing. The crust under the load is faulted down into the mantle and sinks until isostatic equilibrium is attained. The load is, therefore, isostatically compensated by "local compensation" (ideal isostasy). There is no isostatic anomaly. A negative Bouguer anomaly is present only in the area of down faulting.

The type of crustal equilibrium in a given area depends upon the condition of loading. Very local topographical differences due to valleys and hills may produce loads that are easily supported by the crust. Thus conditions of mechanical equilibrium may exist locally. Widely distributed loads of relatively light material such as the water of Lake Bonneville and of Lake Mead, and the Pleistocene ice sheet over Fennoscandia and over the Great Lakes region, may have distorted the crust elastically. There the crust would be in a state of isobaric equilibrium. Finally, concentrated heavy loads of some critical size, such as mountain ranges (e.g., Sierra Nevada), are supported by mountain roots and are probably in a state of isostatic equilibrium.

The type of crustal equilibrium depends also on the prevailing stress conditions. At times of tectonic disturbance, gravity faulting along large crustal faults cannot occur when the tangential pressure exceeds the radial pressure. Isostatic adjustment along crustal faults would, therefore, be pre-

vented. For example, the island arc regions have just undergone, or they may still be in the dying phases of, tectonic disturbances; the crust probably is in a state of isobaric equilibrium (Gunn, 1949; Officer et al., 1957), as shown by figure 3a. At times of relative quiescence, gravity faulting along large crustal faults can occur when the tangential pressure no longer exceeds the radial pressure. Isostatic adjustment would then be possible (fig. 3b). There is evidence that isostatic adjustment is either taking place or that the equilibrium is partially restored in the island arc regions today (Ewing and Heezen, 1955; Officer et al., 1957; Vening Meinesz, 1954; Kumagai, 1940).

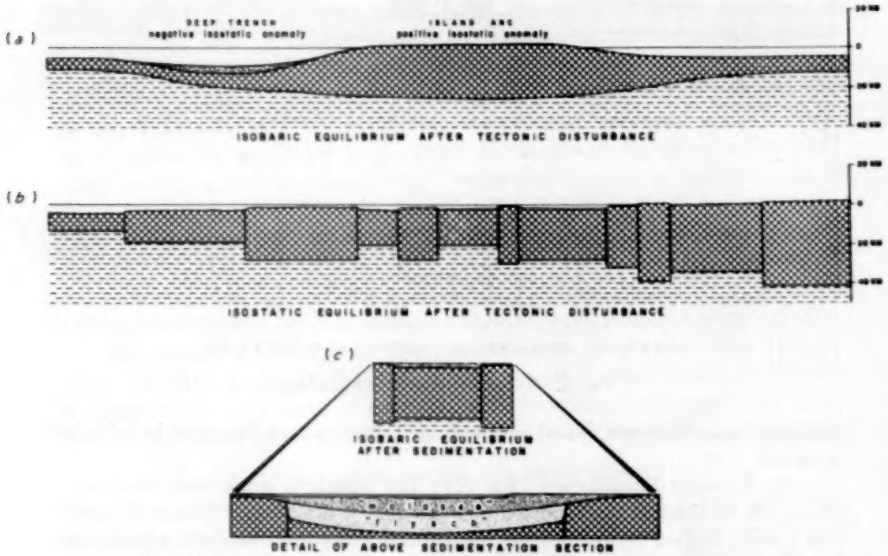

Fig. 3. Tectonic disturbance, sedimentation and crustal equilibrium.

Tectonic Disturbance, Orogeny, and a Theory for the Origin of Geosynclines.—Orogeny as originally defined by G. K. Gilbert (1890, p. 340) is simply "the process of mountain formation," and "the displacements of the earth's crust which produce mountain ridges are called *orogenic*." In particular Gilbert was referring to the process which caused the uplift of the Wasatch mountains of the Basin and Range Province. Subsequently, the term orogeny has been applied to designate structure-transforming movements or large-scale horizontal movements characterized by folding, thrusting, and other structure changes (Stille, 1924; Longwell, 1940). However, folding and thrusting movements are not necessarily mountain forming (Berry, 1929), and mountain chains could be formed by isostatic adjustment after folding (Reid, 1922), i.e., by vertical displacements of the earth's crust and processes that do not involve large-scale horizontal movements. Thus, there is a need to distinguish between the process of mountain forming by horizontal movements and the process by vertical uplifts. The term *tectonic*

disturbance is used here to denote processes which tectonically upset the equilibrium of the earth's crust, such as large-scale folding and thrusting. Such processes are thus anti-isostatic. The term *orogeny* is used here, as originally defined, to denote processes of mountain forming. Tectonic disturbances are commonly orogenic; such processes are designated herein as *anti-isostatic orogenic movements*. On the other hand, isostatic adjustments after a tectonic disturbance may also form mountains; these processes are designated here as *isostatic orogenic movements*.

Discussions on orogeny in geologic time have not always distinguished the anti-isostatic orogenic movements from the isostatic orogenic movements. However, these two processes are genetically distinct processes. Their duration, extent, and periodicity are probably different. Tectonic disturbances, or anti-isostatic orogenies are indicated by prominent angular unconformities and by evidence that the isostatic equilibrium of the crust has been disturbed. Vertical uplifting movements, or isostatic orogenies are indicated by minor angular unconformities, particularly on margins of geosynclines, and by the presence of thick clastic sediments in a sedimentary sequence ("clastic wedge").

Emphasizing the importance of marginal deformation and clastic wedge, Gilluly (1949) has presented a strong case that isostatic orogenic movements at least are not episodic. However, even the most serious critic of the theory of periodic diastophism would not contend that intense crustal deformation is ubiquitous. As summarized by King (1955, p. 737):

> It seems clear that the major orogenic and epeirogenic structures have not been created during relatively short periods of movements, but were built up by movements extending over a long period. It appears unlikely that these movements were continuous; more probably they were episodic, and separated by periods of crustal quiescence. The record, where evidence is relatively complete, suggests that individual orogenic episodes affected relatively small areas; their synchroneity with movements elsewhere is at least difficult to prove by paleontological or other available means of dating.

Episodic (or periodic) tectonic disturbances might be local, yet isostatic equilibrium over wider regions could be upset because of the transfer of crustal materials during such tectonic disturbances. When periods of relative quiescence again prevail in the region, the body forces due to the gravitational field of the earth will tend to restore the isostatic equilibrium of the crust. Regions of thickened crust will form elevations, while regions of thinned crust will form earth depressions. It is here suggested that sedimentation in these isostatically adjusting or adjusted[3] depressions results in the formation of geosynclines.

SEDIMENTS AND STRUCTURE OF GEOSYNCLINES

Thickness of Geosynclinal Sediments and Isostasy.—If geosynclines are isostatically adjusting troughs, the maximum thicknesses of sediments that can be accumulated in them before final isostatic equilibrium is reached de-

[3] The term *isostatically adjusting depression* is used here to denote an area which is isostatically adjusting after the equilibrium of the crust underlying the area is disturbed by a tectonic disturbance. The term *isostatically adjusted depression* is used here to denote an area which is isostatically adjusted after a disturbance so that the area would not subside unless the isostatic equilibrium is again disturbed by sedimentary or other load.

pend upon the thicknesses of the crustal segments at the time that equilibrium was disturbed by tectonic disturbances. Assuming a "standard crust" with a density of 2.84, and a thickness of $T = 33$ km for isostatically adjusted crustal blocks with their tops at sea level, and assuming that geosynclinal sediments do not accumulate above sea level, one can compute the maximum thickness of sedimentary fill in various types of geosynclines from the relation

$$T_c\rho_c = T_s\rho_s + T'_c\rho_c + T_m\rho_m$$

where

ρ_s = density of sediment, or sediment and volcanics
= 2.3, 2.4, or 2.5

ρ_c = density of crust under sedimentary fill = 2.84

ρ_m = density of mantle = 3.27

T_s = thickness of sedimentary (and volcanic) fill $< T_c$

T_c = thickness of crustal block at sea level = 33 km

T'_c = thickness of crustal block under geosynclines $< T_c$

T_m = thickness of mantle which compensates for deficiency in mass of the sedimentary fill
$= 33 - T'_c - T_s$.

The results are tabulated in table 1.

TABLE 1
Thickness of Geosynclinal Sediments

Thickness of Crust (in km) $\rho = 2.84$	Depth of Water* (in km) $\rho = 1.03$	Thickness of Sediments (in km) for aver. density $\rho = 2.3$	Thickness of Sediments and Volcanics (in km)	
			for aver. density $\rho = 2.4$	for aver. density $\rho = 2.5$
5	6.3	12.4	13.8	15.6
10	4.4	10.2	11.4	12.8
15	3.5	8.0	8.9	10.1
20	2.5	5.8	6.4	7.3
25	1.54	3.55	3.95	4.5
30	0.58	1.33	1.48	1.68
32	0.19	0.44	0.49	0.56
32.5	0.10	0.22	0.25	0.28
33	0	0	0	0

* Depth of water in earth depressions had there been no sediments in those depressions.

Sediments (and volcanics) laid down on outer continental margins with a crust 5 to 10 km thick (eugeosynclines) could be accumulated into a conformable sequence up to 10 or 15 km thick. A typical example is the Franciscan group of the Coast Ranges of California, which has an estimated thickness of over 25,000 feet (or 8+ km) (Taliaferro, 1943). The Keewatin and Timiskaming rocks of the Precambrian shield may be another example (Pettijohn, 1943). The thinness of the crust under those geosynclinal regions

is also suggested by the abundance of basic and ultrabasic igneous rocks interlayered with the sediments.

A conformable sequence 6 or 8 km thick could accumulate on an inner continental margin (continental borderland) with a crust 15 to 20 km thick (epi-eugeosynclines), e.g., the Pliocene sediments of the Ventura basin, up to 20,000 feet thick (Natland and Rothwell, 1954). The Stanley-Jackfork-Atoka sequence of the Ouachitas (van der Gracht, 1931; Pettijohn, 1957) is another example.

Sediments laid down on the continental crust 20 to 30 km thick near continental margins (exogeosynclines) could form a conformable sequence about 2 to 5 km thick. The Martinsburg-Juniata-Oriskany and the Romney-Jennings-Catskill sequences of the central Appalachians are typical examples (Pettijohn, 1957).

Sediments laid down in intracratonic basins (intracrational geosynclines), below which the crustal thickness is about 30 km, could form a conformable sequence of about 1.5 km, e.g., the Silurian sediments (4500 feet thick) of the Michigan basin (Kay, 1951).

Finally, sediments deposited on a stable shield area should be, and are commonly observed to be, very thin, not over a thousand feet.

The Triassic rocks of the Atlantic Coast have a maximum thickness of about 25,000 feet or more (Kay, 1951). According to table 1, the crustal thickness should be about 15 km. Or, if it is assumed that the Triassic sediments were deposited in an intermountain basin which had a final elevation of about 1 km above sea level, and that the average density of these sediments and volcanics was 2.4 or 2.5, the crust beneath a Triassic trough containing 25,000 feet (8 km) sediments could be as thick as 25 km.

Thus, the predicted thickness is in general agreement with geological observations: eugeosynclines on continental margins are sites of thickest sediment (and volcanic) accumulations(whereas cratonic shield areas are barely covered by sediments.

Nature of Sediments in Isostastically Adjusted Depressions.—The original depth of water of an isostatically adjusted depression before sedimentation depends on the thickness and density of the crustal column below. Assuming a crustal density of 2.84 and a mantle density of 3.27, and assuming again that the crustal block with its top at sea level is 33 km thick, one can tabulate the hypothetical original depths of water of isostatically adjusted blocks of different crustal thickness (table 1).

Selected geologic examples give some support to these deductions. Eugeosynclines, located along outer continental margins, such as the late Mesozoic Franciscan of the Coast Ranges, characteristically include a thick sequence of deep-water clastics (graywacke and abyssal clays), submarine igneous rocks (ophilolites), and radiolarite (probably an abyssal pelagic deposit) (Steinmann, 1926; Bailey, 1936). The late Tertiary Ventura basin, an epi-eugeosyncline, located on the inner continental margin, received mostly bathyal sediments (flysch) until Pleistocene time when shallow marine and continental sediments (molasse) were deposited (Natland and Rothwell, 1954). Thinner flysch and thicker molasse sediments of the central Appalachians (Pettijohn,

1957, p. 641) may represent deposition in areas near continental margins. Sediments in intracratonic basins are commonly deposited in shallow waters; although supposedly deep-water sediments (deposited below wave base) are also known (Mississippian of Ohio, Rich and Wilson, 1950). That shield sediments are deposited almost universally in shallow water has long been recognized.

The idea that geosynclinal sediments are deposited in isostatically adjusted depressions implies that thick geosynclinal sequences necessarily start with deep-sea and end with shallow marine or alluvial sedimentation (Lawson, 1942). The recognition of flysch and molasse sequences in many geosynclinal sequences, (van der Gracht, 1931; Pettijohn, 1957) lends support to the idea. On the other hand, some geosynclinal sequences, such as the Cambro-Ordovician limestones of the Appalachians, apparently did not start with deep-water sedimentation. In other instances, such as the Triassic of Connecticut Valley, geosynclinal sedimentation definitely did not start under deep marine conditions; taphrogeosynclinal sediments 10,000 to 20,000 feet thick are often continental throughout. Furthermore, a close examination of the stratigraphic record (see fig. 5) will show that the probable water depths at the beginning of most geosynclinal sequences is not so great as predicted by the hypothesis (table 1). Thus, we are forced to conclude that the Lawson hypothesis of geosynclinal subsidence due to sedimentary load in isostatically adjusted depressions must be modified.

Rate of Sedimentation, Rate of Subsidence, and Geosynclinal Cycles.—The fixed relationship between the initial depth of water and final sediment thickness postulated by Lawson (1942) is based on the assumption that geosynclinal sediments are deposited in isostatically adjusted earth depressions. This fixed relationship does not hold if one accepts the geosynclinal theory here presented that geosynclinal subsidence has been initiated by isostatic adjustment following tectonic disturbance whereas sedimentary load has only contributed to further depression of the crust by isobaric adjustment. Then depth of water of the subsiding depressions depends upon

1. the elevation of the depression (with respect to sea level) immediately after the tectonic disturbances which had disturbed the crustal equilibrium of the region.
2. the rate of subsidence of the isostatically adjusting geosyncline.
3. the rate of sedimentation in the isostatically adjusting geosyncline.

The rate of isostatic adjustment by crustal faulting can be estimated[4] if one assumes that vertical movements along steeply dipping faults represent such an adjustment after an anti-isostatic tectonic disturbance. For example, the coastal terraces of southern California have been uplifted since the isostatic equilibrium of the region was disturbed by the Mid-Pleistocene tectonism. The coastal terraces of Ventura County have been uplifted 1250 feet or so since the medial Pleistocene time about 500,000 years ago (Putnam,

[4] The rate of isobaric adjustment can also be estimated. The maximum rate of isobaric uplift in Fennoscandia is probably over 10,000 meters or 30,000 feet per million years (Daly, 1941, p. 309-326). The Boulder Canyon region has subsided at a similar rate under the load of water in Lake Mead (Thomas, 1954). Such rates are far greater than that of sedimentation in any known geosyncline (Kay, 1955).

1942) giving an average of about 2500 feet per million years as the rate of isostatic adjustment. The rate of subsidence of a presumed isostatically adjusting basin can also be estimated directly on the basis of sedimentary history. The floor of the Ventura basin, for example, subsided almost 10,000 feet during the 5-million-year interval between late Miocene (Mohnian) and early Pliocene (Repettian), although the basin itself was never that deep because of contemporaneous sedimentary fill (see fig. 5). The probable rate of isostatic subsidence was, therefore, about 2000 feet per million years.

The rate of sedimentation of a basin depends upon the availability of sediments and upon the rate of subsidence. The average rate of sedimentation in geosynclines has been estimated at from less than 100 to over 3000 feet per million years (Kay, 1955). The average rate for most geosynclines is probably about 500 to 1000 feet per million years, definitely less than the rate for isostatic adjustment here estimated. Therefore, a deepening of an isostatically adjusting basin should take place. After a basin is isostatically adjusted, the average rate of subsidence under sedimentary load depends upon, and is necessarily less than, the average rate of sedimentation. The later geosynclinal history should, therefore, be one of sedimentary fill. Finally, compaction of sediment should cause further slow subsidence of the surface of deposition to allow the deposition of shallow marine and deltaic sediments, and the

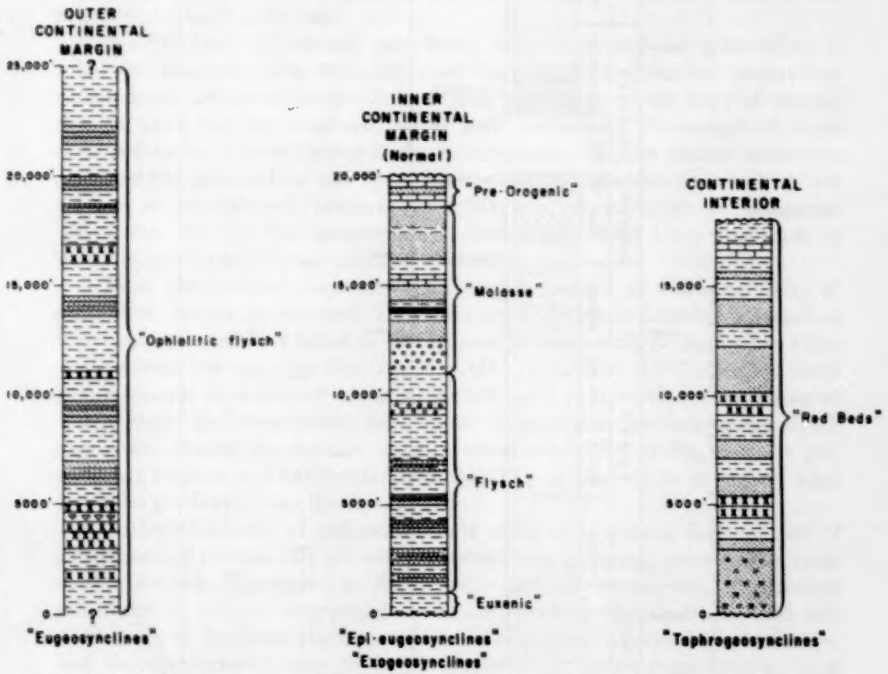

Fig. 4. Geosynclinal cycles.

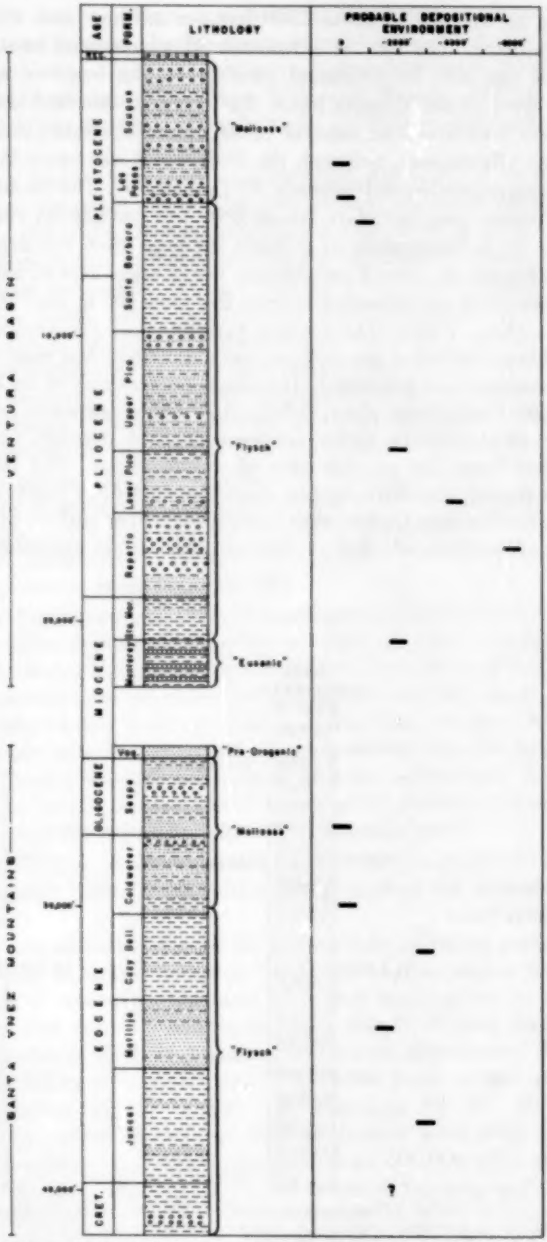

Fig. 5. Geosynclinal cycles of southern California (modified after Natland and Rothwell, 1954).

weight of those sediments should cause further isostatic subsidence. From this deductive reasoning a normal geosynclinal cycle should be characterized by the following stages:

1. Isostatic adjustment after tectonic disturbances results in a rapid subsidence of geosynclinal basins. Accumulation of pelagic and chemical euxinic sediments in the subsiding basin cannot keep pace with the subsidence. The basin is "starved" and often subsides from near sea level to bathyal depth.

2. Isostatic anomalies produced by tectonic disturbance are compensated by isostatic adjustment. However, continued sedimentation prompts continued isobaric adjustment by sedimentary load. Meanwhile, isostatic uplifting of mountains surrounding the isostatically subsiding basin results in conditions favorable for turbidity current deposition. Slow deposition of pelagic muds is complemented by rapid deposition of sediments of turbidity current origin, and the flysch sequence quickly fills the basin.

3. The load of flysch sediments induces quick isobaric subsidence and the compaction of underlying sediments induces slow surface subsidence to allow for the accumulation of deltaic and alluvial molasse sediments. The load and compaction of the accumulating molasse induce further subsidence and more molasse sedimentation. The isostatic equilibrium is maintained, because negative isostatic anomalies on delta flanks are compensated when those flanks are filled by later sediments.

4. The limit of isostatic subsidence is reached, and the geosyncline is filled up. However, slow compaction of the molasse and further compaction of the flysch sediments induce further slow subsidence of the floor of deposition to allow for the accumulation of shelf sediments. The weight of those shelf sediments induces further isobaric subsidence. Shallow marine sediments overflow the geosyncline and spread into adjoining positive areas which are eroded or peneplaned, forming thin blankets of pre-orogenic sediments (Pettijohn, 1957). The geosyncline is completely filled when the limit of isostatic and compactional subsidence is reached.

Most geosynclinal sequences do record a history of sediment-filling of deep-water basins as deduced. That deep-water flysch is followed by shallow water molasse was first noted in the Alps and subsequently recognized in other geosynclinal systems (van der Gracht, 1931). Pettijohn (1957) further noted that a normal geosynclinal cycle is characterized by an orderly succession of four distinct sediment types: the euxinic dark shales, the deep water flysch sands and shales, the shallow water molasse sands and shales, and the pre-orogenic clastics and carbonates (fig. 4). This normal cycle is exactly what would be predicted from the theory.

If the initial rate of sedimentation is equal to or greater than the rate of subsidence of isostatically adjusting geosynclines, a normal geosynclinal cycle cannot develop. For example, the average rate of deposition of continental sediments in ancient intermountain basins, such as the Devonian Old Red sandstones of Scotland, the Upper Triassic Newark of eastern North America, and the Oligocene Molasse of the Swiss pre-Alps, ranges from 1900 to 2600 feet per million years (Kay, 1955). The average rate is about equal to, and

the initial rate might even be greater than, the initial rate of geosynclinal subsidence estimated here. Under such circumstances, a deep-water basin might not be formed even in regions of thin crust. Assume, for example, that a crustal block 25 km thick was elevated to sea level during tectonic disturbance. The block would subside and be 1.54 km below sea level when the isostatic equilibrium was restored *if* there had been no sediment accumulated in the trough (see table 1). However, if as soon as the subsidence started, sediments were introduced at a rate equal to the rate of subsidence, the sedimentary fill would then keep the surface of the subsiding basin more or less at sea level, and 3.5 to 4.5 km of shallow water or continental sediments would have accumulated in the subsiding basin when the crustal block was finally isostatically adjusted (see table 1). The Triassic sediments of the Atlantic Coast were deposited in intermountain valleys where sediments were readily available from surrounding mountains. The initial rate of sedimentation was equal to, or probably exceeded, the rate of subsidence, so that the valley floors of deposition were probably never much below sea level, and 20,000 feet or more of continental sediments could be deposited. Other taphrogeosynclines may have a similar depositional history. Thus a geosynclinal cycle in the continental interior may consist entirely of continental sediments (see fig. 4).

In regions with a very thin crust (10 to 15 km), a normal geosynclinal cycle may not have an opportunity to be developed completely. For example, the average rates of sedimentation of eugeosynclines range from 500 to 1000 feet per million years, and the initial rate of pelagic sedimentation should be much less. A region with a thin crust and a slow rate of sedimentation should be rapidly deepened into a deep-water basin (if it was not already under deep seas after tectonic disturbance). Because of the thinness of the crust, 10 or 12 km of sediment must be deposited in such a basin before it is shoaled to allow shallow marine sedimentation. However, the crustal equilibrium of a region may again be disturbed before the basin is filled to prevent the completion of a normal geosynclinal cycle. Thus, geosynclines on outer continental margins consist almost entirely of deep-water sediments and submarine volcanics and intrusives (fig. 4).

Primary Structure of Geosynclinal Basins and Isostatic Origin of Geosynclines.—Geosynclines are not simple elongate troughs. Stratigraphic studies indicate that geosynclines, or geosynclinal systems, are composed of a number of depositional basins, each of which has its individual tectonic and sedimentary histories (e.g., Reed, 1933; Kay, 1942). The primary structures of those ancient depositional basins are obscured by later folding and faulting. However, considerable evidence has been presented to indicate that geosynclinal basins are bounded on one or both sides by high-angle faults (Clark, 1932; Ver Wiebe, 1936; Moody and Hill, 1956). The continental borderland off the southern California coast, a remnant of the Mesozoic-Tertiary Coast Range geosynclinal system, is characterized by a basin and range topography suggestive of crustal faulting (Shepard and Emery, 1941). The crust below the unconsolidated and semiconsolidated geosynclinal sediments of the present Atlantic Coast of North America is probably also faulted (Officer and Ewing, 1954). More definite examples of geosynclines or basins

of sedimentation bounded by primary faults are the taphrogeosynclines of the Upper Triassic of eastern North America (Longwell, 1922; Stose, 1949), and the alluvial valleys of the Basin and Range Province.

Analyses of the patterns of tectonic lineament also led to the conclusion that the origin of geosynclines is related to crustal faults (Bucher, 1933, Opinion 11, p. 89-90; Cloos, 1948).

A basin and range structure for geosynclinal systems is consistent with the hypothesis that geosynclines are initiated by isostatic stresses after tectonic disturbance. Step faults bounding mountain ranges have been considered surface manifestations of crustal faults along which isostatic adjustment has taken place consequent to folding and thrusting (Reid, 1922; Lawson, 1936; Griggs, 1939). Whether the basins and ranges so produced are subaerial or submarine depends upon the thickness of the crust. Basins in continental interiors are commonly subaerial and sites for alluvial sedimentation because of the relatively thick crust in these regions (30 to 35 km). Basins along continental margins, on the other hand, should be submarine and sites for deep marine sedimentation because the crust there is relatively thin (20 to 25 km), (see fig. 3b). Whereas the submarine basins of sedimentation may be what have been described as geosynclinal basins, the surrounding "submarine ranges" (offshore banks or islands) may well be what have been described as "geanticlines" or "cordillera" in geologic literature.

Not all geosynclinal sediments are bounded by primary faults. Faulting in the Upper Triassic troughs of eastern North American terminated before sedimentation was complete so that late deposits overlap the boundary fault (Bain, 1932; Stose, 1949). Furthermore, isopach maps indicate that geosynclinal sediments in many regions have been deposited in a down-warped depression (Kay, 1951). According to the theory presented here, faulted and downwarped basins might simply represent two stages of the geosynclinal development. Geosynclines initiated by isostatic readjustment after tectonic disturbance subsided along crustal faults. Further subsidence ensued after those faulted troughs had been filled with sediments. However, the excess weight of the thinly spread sediments is commonly not enough to overcome the cohesion between crustal blocks. The sedimentary load is, therefore, compensated by isobaric subsidence (fig. 3c), causing gentle downwarp of the crust. Onto those downwarped surfaces, younger shallower water sediments were laid down. This deduction is supported by the observation that earlier geosynclinal sediments are commonly "geosyncline-bound" and later geosynclinal sediments "may overflow the geosyncline and spread into adjoining more stable areas" (Pettijohn, 1957, p. 644).

TECTONIC DISTURBANCE AND GEOSYNCLINAL CYCLES

The hypothesis that geosynclinal subsidence has been initiated after tectonic disturbance in order to re-establish isostatic equilibrium necessitates a close time relation between the two events. A cursory examination of the sedimentary histories of the geosynclinal basins in southern California and in the central Appalachians indicates that the origins of those depositional basins are related to major tectonic disturbances in nearby regions.

Geosynclinal Cycles in Southern California.—Three major geosynclinal cycles are tentatively recognized in southern California, and a fourth one has just begun:

1. Late Jurassic (Nevadan) Disturbance and Franciscan Geosyncline. Practically nothing is known about the sedimentary history of the Franciscan rocks of southern California. The Franciscan rocks in the Coast Range to the north are probably deep-water sediments ranging from Late Jurassic to Medial Cretaceous in age (Taliaferro, 1943; Easton and Imlay, 1955; Schlocker et al., 1954; Glaessner, 1949). Meager stratigraphic records suggest that the Franciscan depositional basins started subsiding soon after the "Nevadan Orogeny," which was tentatively dated as Late Jurassic (Taliaferro, 1942; 1943).

2. Mid-Cretaceous Disturbance and Early Tertiary Geosynclines. Stratigraphic columns of the Los Angeles and Ventura basins indicate a Late Cretaceous-early Tertiary geosynclinal cycle in southern California (fig. 5). Sedimentary structures suggest that Upper Cretaceous sediments were deposited in deep waters (Jennings and Troxel, 1954). Conditions for deep-sea sedimentation continued until late Eocene, when shallow marine and continental sediments were deposited (Natland and Rothwell, 1954). Oligocene sediments are mostly continental. Finally, relatively thin but widespread Vaqueros sands were deposited during an early Miocene shallow marine transgression.

The initiation of those Late Cretaceous-early Tertiary depositional basins is probably related to the Mid-Cretaceous "Orogeny" which has been recognized in the Baja California region (Darton, 1921; Woodford and Harris, 1939). That there must have been considerable transfer of crustal materials to upset isostatic equilibrium during Medial Cretaceous is indicated by the radioactive dating of the intrusions of southern California, Sierra Nevada, and Idaho batholiths. The ages of these three botholiths are essentially the same, about 102 million years (Larsen et al., 1954).

3. Mid-Miocene Disturbance and Late Tertiary Geosyncline. The late Tertiary geosynclinal cycle of the Ventura and Los Angeles basins may have started during late Miocene. The water depths in those basins reached their maxima of 4000 to 6000 feet during early Pliocene, and were gradually shoaled by the sedimentary fill during late Pliocene and early Pleistocene (Natland and Rothwell, 1954; also fig. 5).

That the initiation of many new late Tertiary depositional basins in the Pacific Coast region is related to tectonic disturbances during medial Miocene has been generally recognized (Reed, 1933; Reed and Hollister, 1936). Reed (1933, p. 186) stated:

> The change in paleogeographic conditions that allowed this middle Miocene change in type of sedimentation was probably the greatest that took place between the Sierra Nevada disturbance of the Upper Jurassic and the final development of the Coast Ranges as a prominent group of mountains in the Lower Pleistocene.

Crustal equilibrium must have been disturbed by intense [illegible]ding and extensive volcanism during the medial Miocene. When isostatic equilibrium was being restored during late Miocene, the Los Angeles, Ventura, and San

Joaquin basins were deepened by isostatic subsidence, while many intermountain basins were created in the interior of the continent ("Mohavia") to allow for accumulation of thick upper Miocene continental deposits (Hewett, 1955).

4. Mid-Pleistocene "Orogeny" and Recent Geosynclines. Late Tertiary rocks of southern California were folded during the medial Pleistocene (Reed, 1933). Post-tectonic isostatic adjustment is probably now taking place, as indicated by the uplifted Pleistocene terraces. Basins on the offshore continental borderland are probably modern representatives of the Tertiary geosynclinal basins of California. These basins are now receiving deep-water sediments. Meanwhile, basins on land are sites of thick recent alluvial accumulations.

Geosynclinal Cycles in the Central Appalachians.—The earliest geosynclinal cycle in the Appalachians is tentatively recognized in the south. The late Precambrian Occoe series and its equivalent consists of typical flysch sediment (King, 1949; Mellen, 1956). These probable deep-water sediments are succeeded apparently conformably by the Early Cambrian Chilhowee group of molasse type clastics (King, 1949). These flysch and molasse deposits are overlain by Cambro-Ordovician carbonates and orthoquartzites (termed by Pettijohn pre-orogenic sediments), which spread out regionally and are the earliest Paleozoic sediments deposited in the central Appalachians.

The slow subsidence which allowed the slow accumulation of Cambro-Ordovician carbonates may be related to the compaction of the underlying clastic sediments. The Occoe and Chilowee group may have a combined thickness of over 20,000 feet (King, 1949). If it is assumed that a 10-percent reduction of porosity of the clastics due to compaction under load took place during the deposition of the carbonates, there could be a slow subsidence of 2000 feet of the surface of deposition. A total slow subsidence of 8000 feet followed because of isobaric adjustment under the weight of the slowly depositing carbonates (assuming 2.7 for the density of carbonate rocks deposited). In other words, 8000 feet of carbonates could be accumulated slowly in a sediment-compacted and isostatically adjusting trough.

The origination of the Precambrian to Cambro-Ordovician geosynclinal cycle in the southern Appalachians is probably related to late Precambrian tectonic disturbances which were accompanied by igneous intrusions about 600 millions of years ago (Rodgers, 1952).

1. Ordovician Disturbances and Early Paleozoic Geosyncline. The early Paleozoic geosynclinal cycles in the central Appalachians may have started in Middle Ordovician (Trenton) with the deposition of the Martinsburg shale. Primary structures of the Martinsburg sediments suggest their deep-water origin (van Houton, 1954). These flysch type sediments are overlain by the Upper Ordovician Juniata Molasse and by the Silurian and Lower Devonian orthoquartzites and carbonates (Pettijohn, 1957).

The initiation of this early Paleozoic geosynclinal cycle is probably related to tectonic disturbance during Medial Ordovician (Hess, 1955). A Medial Ordovician Blountian disturbance in the southern Appalachians has been postulated from sedimentary criteria (King, 1950). The Blountian dis-

turbance is represented in the central Appalachians by an angular unconformity with discordances up to 25 degrees between the Middle Ordovician Martinsburg (or Jacksonburg) and the Lower Ordovician limestones (Miller, 1939; 1941).

The taconic folding is commonly considered to be Late Ordovician, although it has not been accurately dated (Hsu, 1950).[5] The Middle Ordovician Normanskill shale (Black River or lower Trenton) is folded in New York and is overlain with a marked angular unconformity by the Upper Silurian limestones (Schuchert and Longwell, 1932). This unconformity signifies an episode, or episodes, of folding during the post-Medial Ordovician (after Black River or early Trenton) and pre-Late Silurian interval. In eastern Pennsylvania, the uppermost Ordovician Bald Eagle conglomerate or the lower Silurian Tuscarora (Maysvillian) overlies the Martinsburg with slight angular unconformities of less than 10 degrees of angular discordance (Willard and Cleaves, 1939). However, whether this slight angular discordance in eastern Pennsylvania could be correlated with the intense Taconic folding in New York is questionable. In fact, stratigraphic records suggest that the Medial Ordovician disturbance, represented by the pre-Martinsburg unconformity, was more severe than the Late Ordovician disturbance, represented by the pre-Bald Eagle unconformity. That there was considerable disturbance of the crustal equilibrium is also indicated by the intrusion of serpentines during Medial Ordovician (Hess, 1955). After the Medial Ordovician disturbance, the crustal equilibrium must have been disturbed to such an extent that new geosynclinal basins were initiated in the southern and central Appalachians, and new intracratonic basins in the continental interior (J. L. Wilson, personal communication). On the other hand, there is little evidence that crustal equilibrium was greatly upset during the Late Ordovician disturbance. The early Paleozoic geosynclinal cycle, initiated after the Medial Ordovician disturbance, was not disrupted by Late Ordovician movements; the influx of the Juniata molasse was a natural consequence of the shoaling of the deep-water Martinsburg basin.

2. Acadian Disturbances and Late Paleozoic Geosynclines. Conditions for deep-water sedimentation recurred during early Medial Devonian in parts of the central Appalachians (Rich, 1948; Pettijohn, 1957). In northeastern West Virginia, deposition of euxinic Romney shale on shallow marine Oriskany orthoquartzite marked the initiation of the Devonian geosynclinal cycle. The Romney is overlain by the Jennings flysch and the Catskill molasse (Pettijohn, 1957).

The Mississippian and Pennsylvanian molasse which succeeded the Devonian molasse might represent still another geosynclinal cycle in the central Appalachians which was initiated simultaneously with the deposition of Early Mississippian deep-water sediments in Ohio (Rich and Wilson, 1950). However, the crust of the central Appalachians was so thickened and the rate of sedimentation was so rapid that isostatically adjusting Early Mississippian basins were quickly filled by alluvial and deltaic sediments;

[5] The evidence for the dating of orogenic movements of the Appalachian system has been discussed previously in detail by the writer (Hsu, 1950).

thus the Carboniferous geosynclinal cycle of the central Appalachians was not marked by initial deep-water sedimentation.

The initiation of the late Paleozoic geosynclinal cycles is probably related to tectonic disturbances in the Appalachians commonly known as the "Acadian Orogeny."

An early Medial Devonian disturbance has been recognized in the Chaleur Bay region ("Shickshokian Revolution," Alcock, 1935). Elsewhere, Middle of Upper Devonian rocks lie unconformably on intrusives or on Silurian sediments, and the dating of the "Acadian Orogeny" cannot be accurate (Hsu, 1950).

Late Devonian or Mississippian disturbances are suggested by pre-Bonaventure unconformities in the maritime provinces of Canada (Alcock, 1955).

That the crustal equilibrium was disturbed during Devonian time is also indicated by the abundant granite intrusions in New England which have yielded some Early or Medial Devonian (310 million years, Stieff et al., 1957) and some Late Devonian or Mississippian dates (260 million years, Rodgers, 1952; 230 million years, Stieff et al., 1957).

3. Appalachian Folding and Triassic Geosyncline. The Upper Triassic basins of eastern North America are faulted basins filled with continental deposits 10,000 to 25,000 feet thick. That a normal geosyncline was not developed has been attributed by the writer to the rapid rate of sedimentation of those Triassic red beds.

The relationship between the Upper Triassic basins and the Appalachian folding is not clear, because the date of the folding is not known. If we assume a direct relationship between the two events, the Appalachian folding would have to be Triassic because isostatic adjustment usually took place soon after tectonic disturbances. Evidence for the dating of the Appalachian folding has been reviewed (Hsu, 1950; Spieker, 1956). There is no sound reason why the folding might not have occurred in Triassic time.[6]

The crust in the Appalachian folded belt must have been thickened during the Appalachian folding. The thinning of the crust under the Piedmont belt might be related to removal of surficial rocks by overthrusting, and by erosion during the tectonic disturbance, when isostatic adjustment was prevented by regional tangential compression.

CONCLUSION

To provide a logical interpretation of geodetic and seismic information, it is necessary to assume that the earth's crust is in isostatic equilibrium. Creation of earth depressions by subsidence of the crust is often a necessary response to disturbance of isostatic equilibrium. It seems logical to regard geosynclines simply as isostatically adjusting, or adjusted, earth depressions which have received considerable sediments, so that further isostatic subsidence under sedimentary load can be induced. Thus any continental or oceanic depression can develop into geosyncline, provided that an adequate sediment

[6] Evidence for Triassic deformation in the northeastern Appalachians was recently suggested by Woodward (1957).

source is available. Most geosynclines are underlain by continental crust (Gilluly, 1955) and are often found on continental margins simply because of the presence of adequate sediment supply. Oceanic basins receiving little sediment rarely develop into geosynclines.

In the simplified discussions of this preliminary presentation of a geosynclinal theory, use is made of a crustal model based primarily on Airy's mountain-root hypothesis, and local isostasy is assumed. Future progress in geodetic and seismologic research may provide a better crustal model. Some revisions of the quantitative estimates made in this preliminary presentation are, therefore, anticipated. Meanwhile, the validity of the theory can be tested by the rapidly accumulating detailed sedimentologic, stratigraphic, and structural information about the geosynclinal systems of the world.

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RINGOLD FORMATION OF PLEISTOCENE AGE IN TYPE LOCALITY, THE WHITE BLUFFS, WASHINGTON*

R. C. NEWCOMB

ABSTRACT. The type section of the Ringold formation in the White Bluffs of the Columbia River consists of a stratigraphic thickness of about 620 feet of horizontally bedded continental sediments lying between river level and the tops of the bluffs, from about 340 to 960 feet in altitude. The beds are of middle to late Pleistocene age.

The uppermost 505 feet of the type section, between 455 and 960 feet in altitude, is composed largely of lacustrine sand and silt. The lower part, extending upward from river level at 340 feet to the base of the lacustrine deposits, at 455 feet, is composed of a weakly indurated conglomerate member that was deposited by river currents. The conglomerate member extends also below river level, down to an altitude of about 290 feet. It is underlain by 100 to 290 feet of lacustrine silt, clay, and sand, and some gravel beds. This lower lacustrine composite, a part of the Ringold, is commonly called the "blue clays" section and lies, in turn, upon the basalt bedrock, whose surface is somewhat irregular but is near sea level in the central part of a broad syncline.

The succession of lithologic types is believed to indicate two stages in the relative uplift of the drainage rim, the Horse Heaven Ridge, in a total amount of about 1,000 feet. The Horse Heaven Ridge is believed to have been a part of the impounding rim. Its main uplift must have been more or less contemporaneous with the deposition of the lower part of the Ringold formation and must have preceded the deposition of the upper part in middle to late Pleistocene time.

INTRODUCTION

Purpose of the Paper

Certain aspects of the history and character of the Ringold formation have been the subject of intermittent study during the last 80 years. However, published reports on the formation differ as to what types of material compose the formation and as to the significance of these types in the geologic history of the region. This paper was prepared in an effort to clarify some of the previous work and to assist in reaching more acceptable concepts regarding the Ringold. It contains information derived from work done during 1950 and 1951 by the Geological Survey for the Hanford Operations Office of the Atomic Energy Commission.

General Composition and Relations of the Ringold Formation

Overlying the basalt bedrock of the Pasco Basin are various unconsolidated sedimentary deposits consisting largely of silt, sand, gravel, clay, and volcanic ash. An important part of that sedimentary cover is the Ringold formation (Merriam and Buwalda, 1917).

The type section of the Ringold formation crops out between altitudes of 340 and 960 feet in the White Bluffs of the Columbia River. Those are the only strata to which the name Ringold formation has been formally applied (Merriam and Buwalda, 1917). In this paper the Ringold formation is extended to include the subsurface continuation of that material—in places completely down to the underlying basalt bedrock which occurs at about sea level beneath the White Bluffs. The Ringold formation crops out also to the north and east of the White Bluffs. There the slightly eroded upper surface of the Ringold underlies, or closely approaches the plateau surface for many miles—

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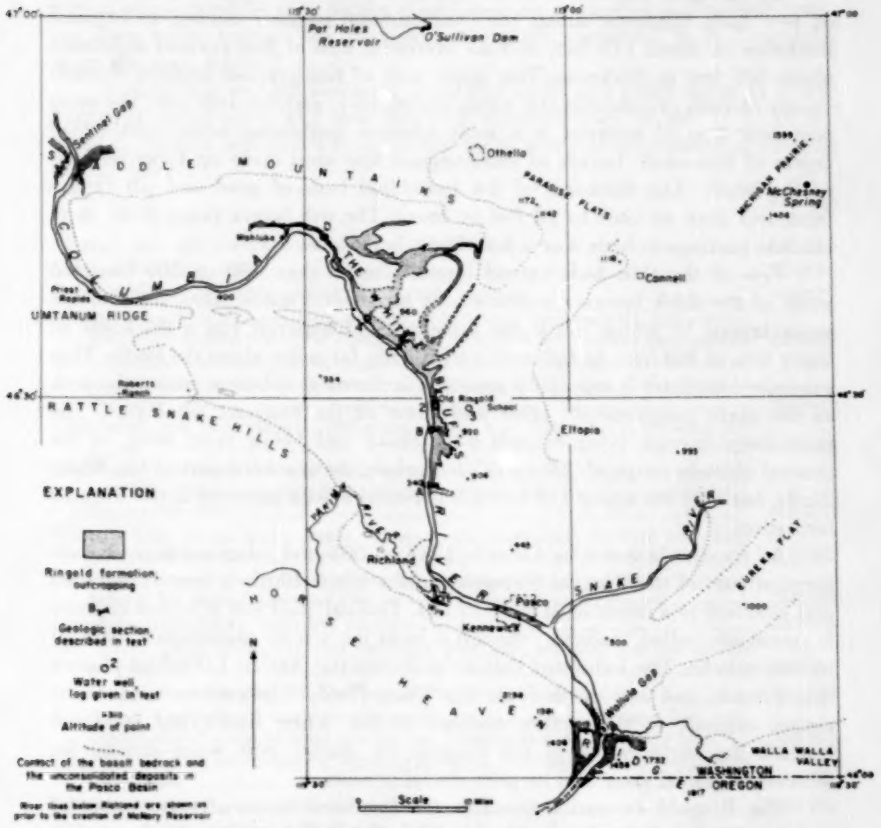
to, and in places beyond, the place where the basalt bedrock emerges at the surface 10 miles north and east of the White Bluffs (fig. 1).


Pot Holes Reservoir
O'Sullivan Dam
Othello
Paradise Flats
McChesney Spring
Connell
Wahluke
Priest Rapids
UMTANUM RIDGE
RATTLE SNAKE HILLS
Ringold School
Ellensburg
Ellensburg
Ellensburg

STRATIGRAPHIC FEATURES OF THE RINGOLD FORMATION

Character and Extent of the Strata in the Type Locality

The Ringold formation in the southern part of its type locality consists of two main lithologic units: the conglomerate member having an exposed thickness of about 115 feet, and an overlying unit of fine-grained sediments about 505 feet in thickness. This upper unit of fine-grained deposits consists mostly of beds of coherent silt, sand, gravel, clay, and volcanic ash. The most prevalent type of material is a weak siltstone containing some interbedded layers of fine sand. Layers of semicompact fine sand make up large sections of the bluffs. The thickness of the individual beds of sand and silt ranges from less than an inch to 10 feet or more. The ash layers range from mere laminar partings to beds 3 or 4 feet thick.

Few of the thin beds extend laterally more than 400 or 500 feet, but some of the thick layers are continuous for several miles. Also, some of the gross layers, in which much the same type of material has a thickness of many tens of feet, can be followed horizontally for miles along the bluffs. That extensive character is especially apparent in the more coherent materials, such as the main conglomerate layer and some of the compact silt layers. The main conglomerate layer extends both above and below river level, in the general altitude range of 290 to 455 feet, along the southern part of the White Bluffs, but only the upper 115 feet has heretofore been included in the Ringold formation.

As previously noted by Culver (1937, p. 60) and other workers, the uppermost part of the Ringold formation in the White Bluffs is heavily calcified and silicified to a depth of at least 15 feet. This calcified and silicified caprock is commonly called "caliche," though it lacks the nitrate constituents inherent in true caliche. The indurated caliche underlies the 900- to 1,000-foot plateau that extends east and north from the White Bluffs. The caliche forms a resistant caprock to the section exposed in the White Bluffs, but is absent beneath the surfaces cut in the Ringold by glacial melt water during the Wisconsin glacial stage and by post-glacial erosion.

The Ringold formation contains the fossilized bones of many types of vertebrate animals and some scattered carbonized wood and other plant matter. On the basis of the vertebrate fossils, the age of the beds was determined as Pleistocene by Merriam and Buwalda (1917), and the fossils were further determined as of middle to late Pleistocene age by Jean Hough (Strand and Hough, 1952).

Material similar in lithology to the type Ringold formation extends downward to the basalt bedrock—or to a thin transitional deposit that may be a pre-Ringold weathering, or soil, zone at the top of the basalt. Although no paleontologic evidence has been secured to prove the material below river level as an extension of the type Ringold formation, the lithologic similarity and the stratigraphic continuity of those "below-river-level beds" are believed to establish them definitely as a downward extension of the Ringold formation of the type locality.

Sratigraphic Sections of the Ringold Formation in the White Bluffs

Geologic sections that were examined vertically across the strata exposed in the White Bluffs are described below; each section was measured with a spirit level from a datum established by controlled barometric leveling from bench marks. The geologic sections are numbered A, B, C, and D, as on the map, figure 1.

A. Geologic section westward down the bluffs at a point north of Pasco Farms pumping plant and near the west quarter-corner of sec. 1, T. 10 N., R. 28 E. Section downward from leveled mark at top of escarpment at altitude of 882 feet:

	Altitude (feet)
Covered, soil, and reworked rubble as of 694-629-ft zone	882-714
Covered, tan, silty soil	714-694
Covered, reworked rubble eroded from caliche caprock	694-629
Siltstone, sandy and clayey, tan, massive at base and progressively more finely laminated toward top; thin sandy layers contain a few pebbles and cobbles	629-623
Sand, medium and fine grained, progressively more silty upward and indurated to sandstone at top, some reworked fragments of siltstone of the Ringold included	623-615
Clay, sandy and silty, massive, buff-green at base and darkening to brown-green at top	615-601
Sandstone, indurated, fine and medium grained, composed mainly of sand-size ash, has minute cross bedding laminae, a strong bed extending horizontally for several miles	601-595
Volcanic ash, clayey, white, dense, massive with conchoidal fracture near base and finely laminated; top part is made up of loose silt-sized shards	595-589
Clay, silty, brown, massive at base but more laminar toward top, in places contains a 2-ft bed of fine sand that is weakly coherent in the center	589-580
Claystone, dark brown, crumbly, massive	580-575
Silt, clayey, grading to silty clay in upper part, tan, massive, top foot is calcareous clay	575-557
Sandstone, similar to 550-542 zone but of whitish-tan color	557-550
Sandstone, fine-grained, weakly indurated, silty and clayey, massive but with highly uniform size of sand grains; top 5 inches is hard cemented layer	550-542
Clay and silt, brown, compact, semiplastic when wet; sandy beds irregularly distributed	542-504
Siltstone, massive, gray, partly indurated	504-494
Silstone, clayey, massive, yellowish at base and gray above; upper 6 ft has many round concretions up to 10-inch diameter; has calcified zone of 1 to 2 ft thickness at top	494-480
Sand, fine, and sitly sand, fine, laminar bedding; material is gray and contains much mica on bedding planes	480-475
Covered, soft material	475-470
Conglomerate, cobble and pebble gravel with some boulders up to 8-inch diameter and with a matrix of quartzose and arkosic medium-grained, well-sorted sand that comprises about one-half of the material; the exotic pebbles and cobbles are mostly quartzite and dense porphyry, granitic types are rare; exotic rock types and Columbia River basalt about equal; locally, beds up to 6 inches thick are entirely sand, firmly cemented, and stand out as resistant ledges; in 435-415-foot zone many pieces of water-rounded and reworked siltstone of the Ringold are present; pebbles are mostly fresh and clean but those of Columbia River basalt have 1/16-inch whitish weathering rind and some granitic pebbles are crumbly	470-415
Covered	415-349

B. Westward down the bluffs at power line in sec. 12, T. 11 N., R. 28 E., from altitude of 716 feet:

	Altitude (feet)
Covered, soil	716-650
Clay, progressively more silty upward, massive, vertical dikes of sand from the sand layer below	650-639
Sand, fine-grained, angular, micaceous, with horizontal and crossbedding, loosely compacted in upper 5 ft; silty sand layers intermixed, basal 3 inches is strongly cemented and contains scattered pebbles	639-606
Clay, silty, tan	606-604
Clay, greenish	604-600
Clay, tan	600-597
Clay, plastic, greenish-brown	597-591
Clay, plastic, tan, iron-streaked	591-588
Silt, massive, tan and yellow banded, becomes more clayey upward	588-567
Sand, silty, fine to medium grained, compact, angular, crossbedded, with concretionary lumps and nodules; upper 10 ft is finer grained, silty, and contains no concretions	567-532
Clay, plastic, jointed, greenish-brown	532-529
Sand, silty, angular, medium-grained, becoming more silty upward; a 6-inch bed at base is iron-stained flaggy sandstone containing a few pebbles and cobbles up to 4 inches in diameter	529-503
Sand, silty, generally loose but compact in places, crossbedded, micaceous, becoming more silty and clayey toward top; light gray	503-490
Clay, silty, tan in lower part grading upward into light brown, plastic, relatively pure clay	490-478
Clay, silty, massive, grading upward into clayey white volcanic ash	478-475
Clay, massive, light tan but brown in lower 4 ft	475-467
Clay, plastic, dark brown	467-465
Clay, silty, massive, tan	465-462
Sand, clayey, well compacted, rudely stratified and progressively more massive toward top; tan colored; has thin laminae of gray clay and iron-stained bedding planes	462-454
Sand, medium-grained, angular, micaceous, tan colored; loose in lower part but compact and well bedded in upper part; upper surface shows channel erosion with 2 to 4 ft of relief	454-433
Conglomerate, well-rounded pebbles and cobbles with a few boulders in a matrix of medium-grained, indurated, clean, angular, quartz sand; rude foreset bedding stratifications are present and the upper 20 ft contains several 6-inch beds of sand that are cemented to hard resistant rock; pebbles and cobbles consist of 70 percent exotic rock types (45 percent quartzite and 25 percent porphyries and granitics) and 30 percent basalt; the basalt particles have a whitish weathering rind up to 1/8 inch thick	433-346

C. Southward down bluffs to junction of a farm lane with "River Road" in NE¼SE¼ sec. 33, T. 13 N., R. 28 E., from altitude of 908 feet:

	Altitude (feet)
Covered; soil with erratic cobbles and pebbles near base	908-891
Caliche; a calcareous and siliceous lumpy horizontal "caprock"	891-884
Claystone, silty, and clayey siltstone; well laminated in 1/4-inch to 2-inch bands, lower half silty and upper half silty claystone; whitish buff color	884-802
Siltstone, sandy, clayey, irregular in firmness and size of grains; a central reddish concretionary zone is a prominent cliff former	802-786
Claystone, silty, and clayey siltstone, massive, bluish color	786-761
Sand, fine and medium grained, loose, micaceous in the partings, intricately crossbedded, top is an erosional surface with 5 ft of relief in 50 ft distance but of a general even altitude	761-741
Siltstone, sandy with many sand partings (cemented) and many concretions; upper and weaker part of a strong siltstone cliff-forming zone whose material is a gradational phase between claystone below and sand above	741-702

Siltstone, bedded in laminae 1/2 inch to 2 inches thick, some clay silt and very fine sand laminae, lowest 4 ft and top part are gradational to material below and above 702-673
Claystone, laminated very finely at base and progressively coarser above; smooth, novaculitelike material, whitish buff 673-662
Sand, fine-grained, firm, gray and brown 662-649
Silt and sand interlaminated 649-640
Sand, silty, medium to fine-grained, rather loose except for 2- to 6-inch cemented layer at base; finely crossbedded, gray sand with silty laminae 640-616
Siltstone, laminated in bands, each 1/4- to 1/8-inch thick, whitish buff, similar to underlying claystone 616-602
Claystone, banded in yellow and gray-buff laminae 1/4 to 1/2 inch thick; a strong cliff former and distinct marker zone of considerable lateral extent .. 602-589
Siltstone, clayey, massive, buff toward base and bluish above, has much nodular concretionary matter 589-582
Sand, medium-grained, with silt interlayered in 3- to 12-inch beds at base but progressively more silty toward top; the sand is finely crossbedded in waterlaid (lake-current) type of deposition 582-570
Sand, medium-grained, loose, clean, cross-bedded, micaceous 570-543
Claystone, silty, dark, damp, with much nodular concretionary matter 543-526
Clay, silty, and clayey silt, blocky and massive, whitish buff in lower 30 ft and dark brown-green toward top; fossil bones of peccary *Platagonus* at 525-ft level 526-471
Clay, silty, dark green and brown but whiter toward top; contains much secondary gypsum, iron oxide, amber and carbonized wood 471-461
Sand, medium- to fine-grained, whitish, with secondary gypsum crystals common at the base 461-458
Silt, partly indurated, whitish 458-454
Silt, sandy toward base but more clayey and more compact upward; top 10 ft is a green-drab claystone that is hard and "flinty" 454-428
Sand, medium and fine-grained, fairly loose, brown; base not exposed 428-426
Covered 426-410

D. Westward down the bluffs to river's edge just east of the north quarter-corner of sec. 17, T. 14 N., R. 27 E., from an altitude of 584 feet. The top is an eroded surface overlain by glaciofluviatile deposits.

	Altitude (feet)
Siltstone, massive, blocky with interlayers of sandy silt laminae and sand	584-547
Clay, silty, like 533-518-ft zone but with more siliceous nodules	547-535
Volcanic ash, white, glass shards of silt and sand (minor) size	535-533
Clay, cream colored; contains silica nodules up to 4 inches in diameter	533-518
Siltstone, clayey, cream to white color, iron-stained	518-488
Claystone, silty, brownish-buff color	488-470
Siltstone, slightly micaceous, massive, gray to white; erodes to a coarse badland-type topography	470-430
Clay, massive, jointed in blocks; contains white calcareous nodules; gray to brown in color	430-418
Clay, silty; in lowest 6 ft has pods and streaks of calcareous deposit but grades upward into darker greenish and blue-gray colored pure clay; upper most 1½ ft are a silty resistant bed	418-409
Siltstone, similar to underlying zone but of darker color and finer grain; joint planes contain calcareous and siliceous impregnation that surrounds globular masses of the siltstone	409-390
Siltstone, slightly sandy, brown; upper 10 ft is rendered resistant by siliceous impregnation; base not exposed	390-373

Drillers' Logs of Wells Through the Ringold Formation

The logs of four wells give some record of the extension of the Ringold formation below river level, as it is penetrated in wells along the Columbia

River near the type locality (fig. 1). The logs are well-drillers' records with stratigraphic designations by the writer.

1. M. C. Waters. Located in SW¼SE¼ sec. 24, T. 9 N., R. 28 E., northeast of the Richland "Wye" at an altitude of about 340 ft

Materials	Thickness (feet)	Depth (feet)
Alluvium, glaciofluviatile(?) deposits and Ringold(?) formation:		
Gravel and sand, gravel and clay	50	50
Ringold formation:		
Clay, blue	72	122
Columbia River basalt:		
Basalt, water-bearing	1	123

2. U. S. Government. Former irrigation well near old Ringold School. Located in SW¼SW¼ sec. 24, T. 12 N., R. 28 E., at an altitude of about 396 ft

Alluvium:		
Sand	12	12
Ringold formation:		
Gravel, cemented	73	85
Clay	20	105
Clay and gravel	22	127
Gravel	28	155
Boulders and gravel	6	161
Clay, blue and black	35	196
Sand	9	205
Columbia River basalt:		
Rock, black	29	234
Rock, black, and shale	6	240
Shale	17	257
Basalt, dark	214	471
Clay, gravel, and shale	25	496
Basalt	43	539
Shale	25	564
Basalt	128	692
Shale and sandstone	28	720
Basalt	35	755

3. U. S. Government. Located in SE¼NW¼ sec. 28, T. 12 N., R. 29 E., at an altitude of about 914 ft

Materials	Thickness (feet)	Depth (feet)
Eolian deposits and glaciofluviatile erratic material:		
Soil, sand, gravel, and boulders	5	5
Glaciofluviatile erratic material and caliche-impregnated Ringold formation:		
Gravel and boulders	9	14
Ringold formation:		
"Sandrock"	76	90
Clay	60	150
Sand, reddish	6	156
"Sandrock"	44	200
Clay	50	250
Sand, gray	10	260

Clay	60	320
"Sandrock"	22	342
Clay	18	360
"Sandrock," blue	40	400
Gravel	5	405
Clay	45	450
Columbia River basalt:		
Basalt, black and brown colored, porous 546-548 ft	140	590
Basalt, brown, "burnt"	11	601
Basalt, broken	31	632
Basalt, black	65	697

4. U. S. Government (at former Foster Ranch). Located in SW¼SW¼ sec. 13, T. 13, N., R. 27 E., at an altitude of about 410 ft

Materials	Thickness (feet)	Depth (feet)
Soil, silt, clay, sand	4	4
Landslide (largely materials of the Ringold):		
Silt, sandy, with clay	22	26
Sand and silt	22	48
Claystone, silty, tan	15	63
Ringold formation:		
Claystone, silty and sandy, tan	57	120
Sand with silt and clay	25	145
Silt with sand and some pebble gravel	5	150
Sand, medium and fine-grained, with silt	45	195
Silt, sandy, quartzose	10	205
Siltstone and claystone, dark, with concretions	10	215
Sand, medium to fine, gray, quartzose	18	233
Siltstone and claystone, brown, quartzose, with concretions and a few pebbles	72	305
Sand, coarse and medium, quartzose	5	310
Sand and gravel, quartz sand with basaltic pebbles	5	315
Columbia River basalt:		
Basalt, gray, weathered	4	319
Shale, black, tuffaceous	5	324
Basalt and tuff interlayered	283	607

LITHOLOGIC ASPECTS OF THE FORMATION

Distinctive Lithologic Units of the Ringold Formation

The lowest part of the extended Ringold formation commonly has been logged by well drillers as blue clays, and that name has come into use locally for the blue and green-colored silts and clayey silts that form the lowest 100 to 300 feet of the Ringold formation as herein extended. In most places they extend upward from the basalt bedrock to an altitude of about 290 feet. Drill cuttings reveal the material to be largely silt and to contain considerable sand, presumably as thin interbeds. Some beds of gravel and sand are interbedded within the blue clays and in some places even predominate over the "clays." Apparently the most distinctive feature is the blue or green color, which suggests that the material (at least at some places) has not undergone

oxidation as have the other clays and silts which lie mostly above the regional water table. The zone of the blue clays is well known to drillers, especially in the Richland area and downstream to Kennewick and Pasco, where it occupies much of the saturated zone and, in places, precludes development of ground water from the Ringold.

Above the blue clays in the southern part of the type locality, the conglomerate member extends from a variable altitude of about 290 feet upward for about 165 feet. It is the most distinctive and the most permeable part of the Ringold formation. It occurs in a linear strip, which averages 10 miles wide and about 50 miles long, between Sentinel Gap and Wallula Gap. The northeast margin of the conglomerate member runs southeast about through Wahluke and the old Ringold School, thence southeast toward Pasco. The base of the conglomerate is little, if any, higher near the north end of the White Bluffs than it is at the south end. Toward its northeastern margin the conglomerate member becomes progressively more sandy until it is made up predominantly of sand in the bluff exposures in sec. 36, T. 12 N., R. 28 E., about 5 miles north of the old Ringold School. From that locality the sand phases continues for several miles farther north. Merriam and Buwalda (1917, p. 261) described a section of 503 feet of the formation in the area of this marginal sandy phase of the conglomerate member. Outside the area of the conglomerate train, in the northernmost part of the type locality of the Ringold, the section between altitudes of 290 and 455 feet is occupied by finer grained deposits (see geologic sections C and D) such as occur above and below the conglomerate in the southern part of the White Bluffs.

The type conglomerate is a rather uniform aggregation of well-rounded pebbles and cobbles and some small boulders; the spaces between the pebbles and cobbles are almost completely filled by a matrix of medium to fine subrounded and angular siliceous sand. The pebbles and cobbles are about 65 percent quartzite and other metamorphic, granitic, and porphyritic volcanic rocks of upriver, or exotic, types and 35 percent basaltic rocks similar to the Columbia River basalt. The sand is largely all upriver quartzose material. Sand lenses and beds are common in the conglomerate; lenses of sandy silt are rare.

The position of the conglomerate member, lying in the 290-455-foot altitude zone, caused it to be exposed over a large part of the surface produced by the Columbia River's erosion of the Ringold formation. Thus, beneath much of the lowland strath of the Richland-Pasco-Kennewick area the gravelly glaciofluviatile and fluviatile deposits lie directly upon an erosional surface cut in the conglomerate member of the Ringold formation. In many wells this circumstance makes difficult the location of the actual point of contact between the two gravelly units.

The position, shape, thickness, and lithology of the conglomerate member indicate that it represents a river-laid train of gravel deposited across a basin in which quiet-water sediments had been deposited previously and were to be deposited later. Thus, it testifies to a time in which the impounding rim was still low enough to allow integration of an active river current across the basin of deposition.

The upper 500 feet or so of the Ringold formation, upward from the conglomerate member to the caliche cap, is not a marked lithologic type. It consists of intergraded layers of fine sand, silt, and clay, some zones of which attain considerable thickness. Some of the massive claystones and siltstones of this unit attain thicknesses of 20 to 40 feet in the northern part of the bluffs and are prominent cliff-forming units which can be followed for miles.

In most of the beds of the finer materials, particle sizes grade progressively toward the type of material next above and below. Thus, the successions of silt, fine sand, clay, and volcanic ash exhibit some rhythmic changes in a vertical direction, as though they resulted from deposition by the gradually shifting of currents in a large lake. However, certain grain sizes seem to dominate in the formation. In places, one particular grain size seems to be generally predominant for a considerable distance vertically and laterally, even though material of that grain size interfingers with beds of other grain sizes. Thick sand beds are most common in the southern half of the White Bluffs, massive siltstones predominate farther north, and thick claystones are most common at the extreme northern end of the bluffs.

General Lithology of the Ringold Formation

Some characteristics of the materials are summarized as follows:

Depositional:	
Rock types	Exotic, or upstream, materials predominate, almost exclusively in particles below medium sand size
Grain sizes	Silt and fine sand predominate, many thick and continuous silt and clay strata being present. One large train of cobble and pebble gravel was deposited, with interstitial sand of fine and medium grain sizes
Sorting	Well sorted, in thin beds mostly gradational to each other horizontally and vertically
Shapes of grains	Gravel is well rounded; grains of silt and fine sand are angular
Alteration:	
Rinds	Alteration rinds 1/32 to 1/8 inch thick on basalt pebbles and cobbles of the conglomerate
Induration	Silts and clays compact; gravels and sands compact and contain cemented layers; fairly resistant to wind erosion
Cementation	Generally lacking or weak, but there are strongly cemented thin layers in the conglomerate, concretions in clays, and some strongly cemented sandstone "ribs" in the sand beds
Secondary products	Secondary gypsum; fossil bone, largely petrified; calcified and silicified "caliche" cap on pre-Wisconsin surface

Cessation of Deposition of the Ringold

Apparently, deposition of the fine-grained materials that make up the uppermost part of the Ringold formation came to a rather abrupt end. That conclusion is suggested by the fact that the final depositional surface is preserved almost undissected in large areas and has been destroyed only in the rather narrow strath in which the Ringold is still being removed by the

Columbia River. The postdepositional erosive currents did not swing widely over the top of the Ringold deposits. This observation is based on the absence of current-transported deposits of late Ringold age over broad plateau areas that now represent the top of the Ringold formation. Either a tectonic or an erosional lowering of the bedrock rim occurred, or there was a change in the erosional capacity of the river. This rather sudden change caused the river to entrench itself quickly in the soft materials and to become fixed in a wide gorge cut in the Ringold formation. It has since remained in this gorge except for the aberrant diversions of the later glacial waters. Melt waters of the Wisconsin glacial stage eroded two main channels across the plateau underlain by the Ringold formation and scattered glaciofluviatile deposits and erratics rather widely to a maximum altitude of about 1,150 feet.

That the climate was dry, at least periodically, is indicated by the strong calcium and silica caliche that developed during the interval between the end of Ringold deposition and the deposition of the glaciofluviatile materials.

RELATIONS OF THE RINGOLD FORMATION TO THE STRUCTURAL DEFORMATION OF THE BASALT BEDROCK

In the White Bluffs the Ringold formation is not tectonically deformed. Many of its marker beds can be traced for miles with negligible variations in altitude. The major bedding planes of the type section are generally less than 1° from the horizontal, though dips as great as 2° can be found at the tapering ends of some beds and layers. Merriam and Buwalda (1917, p. 261) found that the strata in the White Bluffs lie parallel to the water surface of the Columbia River and stated that all recognized deformation seemed to be due to landslides.

The fine-grained materials of the basal 300 feet of the formation as extended in this paper attest to an initial lake-type impoundment during which the basin was filled to, or slightly above, what is now the 290-foot altitude and what was then the level of the overflow rim. The 160-foot-thick conglomerate indicates the re-integrated river deposited it in a trainlike strip across the basin during either a slow subsidence of at least part of the basin, or a slow elevation of the controlling rim. Finer grained materials doubtless were deposited under flood-plain conditions on the flanks of the gravel train. The inclusion of some blocks of siltstone of the Ringold in the upper part of the conglomerate indicates that the preconglomerate (or perhaps contemporary) siltstone was in places sufficiently high to be undercut and eroded along the flanks of the current-laid gravel train.

The upper 510 feet of the type section shows that fine-grained sediments again accumulated after the controlling rim had been reestablished so as to impound a lake again. The continuation of lake-type deposition to the top of the type section indicates the lake probably had not been completely filled with sediment when the controlling rim was breached, the lake was drained, and entrenchment of the river in the Ringold deposits began. The cessation of Ringold deposition was relatively sudden. It was followed by a dry climate in which the caliche cap developed on the undissected parts of the deposit. The

caliche cap existed for some time before the reinvigorated dissection and subsequent deposition of the Wisconsin stage of the Pleistocene.

One inference resulting from the available information is that the rim of the basin, the Horse Heaven Ridge, was uplifted in two stages in middle to late Pleistocene. Another is that strata of the Ringold now below river level at the type locality—the blue clays and the lower part of the conglomerate member—should have been deformed in places near the lines of displacement which cut the basalt bedrock along the Horse Heaven uplift and also along the other edges of the Pasco Basin, if the deformation at those places coincided with that in the Horse Heaven uplift. An appropriate degree of deformation is commonly reported for relatively thin layers of the Ringold at such places as the Potholes (O'Sullivan) dam site (Jones, 1946) 24 miles north of the White Bluffs, at the Roberts Ranch in the Dry Creek valley 26 miles west of the White Bluffs, and in ravines 4 miles south of the Richland Wye. As a result of regional tilting and uplift, thin layers of the Ringold formation may be present at such high altitudes as 1,400 feet at McChasney Springs on Michigan Prairie (Lupher and Warren, 1942, p. 879), and 1,200 feet in the Paradise Flats southeast of Othello. The literature contains no description of deposits of an age and lithologic type comparable to those of the Ringold formation in the area south of Wallula Gap and the axis of the Horse Heaven Ridge.

This interpretation—that the earlier Ringold deposition preceded the main uplift of the Horse Heaven Ridge and, possibly, that of other ridges and slopes farther north, and that the later (postconglomerate) deposition followed the uplift—is believed to clarify many aspects of the occurrence and deformation of the Ringold formation observed beyond the type locality by previous workers (Culver, 1937, p. 60; Jones, 1945).

The reconstructed surface of the original basalt rim at Wallula Gap in the Horse Heaven Ridge is now at an altitude of about 1,400 feet. The rim of the next younger gorge through the gap is now at an altitude of about 1,000 feet, roughly the same as that of the undissected surface on the Ringold northeast of the White Bluffs. The rock bench at 1,000 feet altitude probably represents the outlet rim of the basin at the time deposition ceased. The conclusion (with which the present writer agrees in general) that the uplift of the Horse Heaven Ridge at Wallula Gap occurred just prior to and contemporaneously with the deposition of the Ringold formation, was proposed by Warren (1941). It disagrees with Waters' map (1955) and the statement on page 680 of his paper that the Horse Heaven Ridge has been in existence and the Wallula Gap has been used by the Columbia River since late Ellensburg (Pliocene) time.

SUMMARY AND CONCLUSIONS

1. In addition to strata in the type section, the Ringold formation can be extended downward to include similar strata below the level of the Columbia River at the White Bluffs.

2. The materials of the Ringold formation in the type locality are largely derived from upriver sources. The grains below medium sand sizes, with the

exception of small amounts of volcanic ash, are almost entirely from upriver sources.

3. The average section at the White Bluffs includes a blue clays unit of silt, clay, sand, and gravel which extends upward from the basalt bedrock at about sea level to an altitude of about 290 feet, a 160-foot thickness of weakly cohesive gravel and sand called the conglomerate member of the Ringold formation, and an upper unit of silt, sand, clay, volcanic ash, and gravel which lies in the general altitude range of 450 to 960 feet. A 15-foot-thick layer of "caliche" caps the section. It originated prior to the Wisconsin glacial stage.

4. The Ringold strata in the type locality are essentially horizontal but depositional conditions imposed local dips as great as 2 degrees. The strata above the conglomerate member have features that indicate deposition by shifting currents in a lake. The conglomerate member was deposited by river currents.

5. The strata of the Ringold formation indicate a history which includes, successively, (a) impoundment of the Columbia River during the deposition of the blue clays, (b) integration of the course of the Columbia River across the basin and the deposition of a 160-foot-thick gravel and sand train, the conglomerate member, (c) reimpoundment of the river, during which time the upper 510 feet of the deposits of the type section were laid down by shifting currents, as an incomplete fill of a lake basin, (d) drainage of the lake and rapid entrenchment of the river into the Ringold formation after breaching of the bedrock rim at Wallula Gap, and (e) development of a caliche cap layer on the Ringold prior to the arrival of the glaciofluviatile and fluviatile deposits at the time of the Wisconsin glacial stage farther upriver.

6. The controlling agent in the two-stage impoundment of the Columbia River during Ringold time was the uplift of the Horse Heaven Ridge. As the depositions of the Ringold formation and the uplift were contemporaneous, the age of the main uplifts of the Horse Heaven Ridge is interpreted as middle to late Pleistocene.

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LATE-GLACIAL AND POSTGLACIAL VEGETATION FROM GILLIS LAKE IN RICHMOND COUNTY, CAPE BRETON ISLAND, NOVA SCOTIA

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ABSTRACT. Pollen analysis of a deposit from Gillis Lake, Richmond County, Nova Scotia, has yielded a stratigraphy consisting of nine zones, two of which are tentatively divided into sub-zones. The three bottom zones are characterized by the pollen of open vegetation, particularly of heaths, willow, sedges, grasses and *Artemisia*. The second of these three zones contains more shrub pollen than the other two, and appears to represent a transitory warm time corresponding to the Two Creeks interval. The upper limit of these late-glacial zones has been carbon-dated at 10,340 ± 220 years, which suggests that this part of Nova Scotia was deglaciated shortly before Two Creeks time and has been ice-free since. The amount of coniferous pollen in the three late-glacial zones suggests that the vegetation was not completely open, but the presence of tree species immediately after deglaciation has not been definitely established. The vegetation was certainly open enough to permit much wider ranges of a number of plants now restricted to isolated bare-ground habitats, including, by inference, many of the Cordilleran disjunct species. *Ulmus*, a genus whose native status in Nova Scotia has been in doubt, has been part of the vegetation since the entry of mixed deciduous forest, and *Artemisia* have been intermittently present during post-glacial time. The influence of European settlement on the vegetation is clearly shown in the top sub-zone of the profile.

INTRODUCTION

Although eastern Canada was the first part of the North American continent to be investigated pollen analytically (Auer, 1927a; 1927b; 1928; 1930; 1933) very little work has been done on its Pleistocene paleo-ecology in the past twenty-five years. Two unpublished honors theses (Hall, 1949; Killam, 1951) from Acadia University have extended reconaissance studies of the type carried out by Auer to parts of the region he did not visit.

This is regrettable, for the part of Canada that surrounds the Gulf of Saint Lawrence has a flora that is rather interesting. The floristic evidence can be construed to indicate a complex Pleistocene history while the geological evidence favors tabula rasa glaciation late in the Pleistocene, with only a few rather dubious local advances since. Despite the unpromising nature of the geological evidence, the floristic material is so strongly suggestive of varied late-Pleistocene events that the area deserves detailed stratigraphic investigation. A number of the points of dispute among biogeographers, such as the nunatak hypothesis of Fernald (1925), and the forest-free corridor theory of Marie Victorian (1925; 1938) are ones that are quite susceptible to stratigraphic investigation. Indeed, to discuss such ideas without an adequate background of information regarding the number, timing and extent of the ice advances, the nature of the vegetation since deglaciation, and the relative position of land and sea, invites a futile cycle of assertion and counter-assertion.

We are undertaking an extensive study of the Pleistocene history of Nova Scotia, in connection with investigations of lake development, that we hope will shed some light on the biogeographic problems of the area. The first piece of work, a pollen analysis of a lake profile from southern Cape Breton Island, forms the basis of this paper.

The profile is from Gillis Lake in Richmond County, N. S., a senescent pond by the west side of the road between Barra Head and L'Ardoise, at 45° 39′ 30″ N., 60° 46′ 30″ W., lying between the 150- and 200-foot contours. It is a rather shallow body of water, and over most of its area the surface is choked with dense beds of *Juncus* and *Nymphaea odorata*, with smaller amounts of *Scirpus*, *Pontederia cordata*, *Eriocaulon septangulare*, *Lobelia dortmanna* and *Nuphar variegatum*. We took our boring in 3 meters of water in the small vegetation-free eye after some preliminary sounding to find the deepest water.

This lake was chosen because it lies in the part of the Province where the striae and drumlinoidal forms lie in a prevailing northwest-southeast direction, only a few tens of miles south of the area described by Goldthwait (1924) where the striae and other indicators of ice travel suggest a direction of movement to or from the northeast. H. L. Cameron, using air photos, has found that the latter area is bounded on the south by what appears to be a complex set of terminal moraines and has suggested that they mark the limit of advance of the northeastern glacier (personal communication). We chose Gillis Lake as an easily accessible locality which seemed to lie on till of the main continental glaciation, yet was sufficiently close to the area of the Cape Breton County advance to permit correlation with profiles from it.

METHODS

Our field operations were carried out on a plywood platform supported between two rubber dinghies, using a piston sampler of the type described by Vallentyne (1955) embodying an expanding rubber piston more or less of the type described by Rowley and Dahl (1956). Our model had a diameter of two inches and, although it worked well at Gillis Lake, was too large for convenient work in stiff clay. With it we penetrated almost six meters of lake sediment with ease, and entered the underlying drift.

Some pains were taken to obtain a core that was not truncated at the top. A piston sampler is capable of taking a very good undisturbed sample of the mud-surface, provided that the drive is started in the water, as demonstrated by Brown (1956) in modifying the sampler for investigation of the mud-water interface. In our work it was not convenient to maintain the upper sample tube in a vertical position until examination of the sample could be carried out, so we carefully decanted the water and closed the tube with a cork stopper at either end. Pushing in the cork disturbed the sediment for a distance of a few centimeters, and caused some sediment to squirt out, so that a few centimeters were lost from each meter sampled. For this reason we did not actually obtain the surface layer of lake mud, but we came much closer to doing so than is usual in work of this sort.

The samples were collected in the late summer of 1955. During the winter of 1956-7 they were extruded from their metal tubes, a clean surface was prepared on the extruded core, and samples of ⅛ or 1 milliliter were removed at intervals for microscopic examination. These samples were boiled briefly in KOH; strained if necessary through a brass sieve to remove coarse plant fragments and bryozoan statoblasts; run through the bromoform flota-

tion procedure, basically as described by Frey (1951); acetolyzed; boiled briefly in KOH once more and mounted in glycerine with safaranin. Some samples required a second bromoform flotation after acetolysis, and a few did not require bromoform at all, but the procedure described was followed with most of the material.

The point on a piston core sample whose position can be determined with the greatest accuracy is generally the top, and this was used as a datum in measuring off subsampling intervals on each piece of core. Because of losses associated with corking, as mentioned above, the total length of the samples did not come quite to one hundred centimeters for each one-meter piece, and the gap, which was under five centimeters in all cases, was conventionally placed at the bottom of each piece of core.

The uppermost piece, taken in a drive begun in the water, lacks such a datum. The upper sample was allotted an arbitrary gap of five centimeters, and the rest of the difference between the length of this sample and one hundred centimeters was assumed to represent lake water decanted from the tube before corking. This explains why, in the pollen diagram, the distance to the one-meter mark is only about 0.6 of the distance between the other meter marks.

Counting was carried out at 660 times magnification using a Spencer microscope fitted with apochromatic objectives, an achromatic condenser of numerical aperture 1.30 and a microscope lamp arranged to give Koehler illumination. As almost all of the material was mounted in glycerine it was possible to roll the grains freely, a great help in making identifications of collapsed grains, but oil was not used. Careful adjustment of the correction collar on the apochromatic objective provided our highest resolution.

Our reference collection is not quite complete for the Nova Scotia flora, which contains 1323 species (Roland, 1945). We have about 1200 species but some of them are represented by rather poor slides. Most of the species in our collection are native or introduced to the Province, but a sizeable fraction are species of more northerly distribution and a handful are plants of so southerly a range that they can hardly be expected in any postglacial deposit from Nova Scotia.

Before beginning the analysis several months were spent in gaining familiarity with the reference collection, making notes on the species and organizing the notes in the form of an analytical key. During the counting repeated reference was made to the collection in identifying new or unfamiliar grains, but we have no strong feeling of confidence in the identifications. We have indicated our uncertainty by including a question mark after the most dubiously identified categories in the table and the diagram.

ACKNOWLEDGMENTS

We wish to thank Miss Jean Turner for her help in making up a large part of our reference collection, the National Research Council of Canada for a postdoctorate research fellowship held during the field work, and the Nova Scotia Research Foundation for defraying the field expenses and for financial assistance in making up reference slides. Preparation of the fossil material

for microscopic examination was aided by a grant from the American Academy of Arts and Sciences; Miss Evelyn Campbell, librarian of the Nova Scotia Research Foundation, gave us very valuable bibliographic assistance and Dr. E. Gordon Young, Director of the Maritime Regional Laboratory, permitted us to use the machine shop facilities of that laboratory in constructing the boring apparatus.

GROSS STRATIGRAPHY

Gross examination of the core, as well as casual inspection of the gauze strainings under a dissecting microscope, suggests that the entire deposit is of freshwater origin. There is no obvious evidence of an unconformity in the section, and we believe that it represents uninterrupted deposition.

The base of the section sampled in the field consisted of coarse pinkish gravel with included fines of the same color. The coring operation was terminated when we could no longer drive the sampler into this material, presumably because we had struck a piece of stone too large to enter the two-inch sampling head.

As we took the bottom piece of core from the lake a considerable quantity of the basal gravel fell out of the sampler and when we extruded the core in the laboratory we found that none had been collected. The lowest layer remaining consisted of about one centimeter of pink clay grading conformably into the sediment overlying it.

The rest of the bottom meter consisted of a gray clay laminated by indefinite horizontal black bands. These bands were not pronounced enough to be measured or counted, but there appeared to be about 5 to 50 of them in each 10 centimeters. The next meter of core consisted of similar laminated material, but with a slightly different color, having a very definite greenish cast. Material of this sort also composed the base of the fifth one-meter sample, but above this there was a ten-centimeter layer without bands and with a very faint pinkish cast. This was overlain by about seventeen centimeters of banded gray material similar to that which formed the bulk of the bottom meter, which was, in turn, overlain by a greenish gray-brown gyttja. This gyttja extended up into the fourth meter, where it changed to a rich dark brown sediment with occasional scraps or layers of coarse plant detritus that continued to the surface. All of these facies changes were gradual ones, except for the sharp boundaries of plant detritus layers.

Organic matter in the core was estimated by determining the loss of weight on ignition of oven-dried samples, each with an original wet volume of about $\frac{1}{8}$ ml. The resulting curve, together with a synopsis of the gross stratigraphy, can be seen in figure 1. The basal clay, as might be expected, is extremely inorganic and the banded gray lake sediment is low in carbonaceous matter. The greenish-gray lake sediment is somewhat more organic, while the brownish lake sediments lose about fifty percent of their dry weight on ignition.

MICROSCOPIC STRATIGRAPHY

The results of the pollen and spore analysis are summarized in figure 1. In addition, a number of species are to be found in table 1. These were

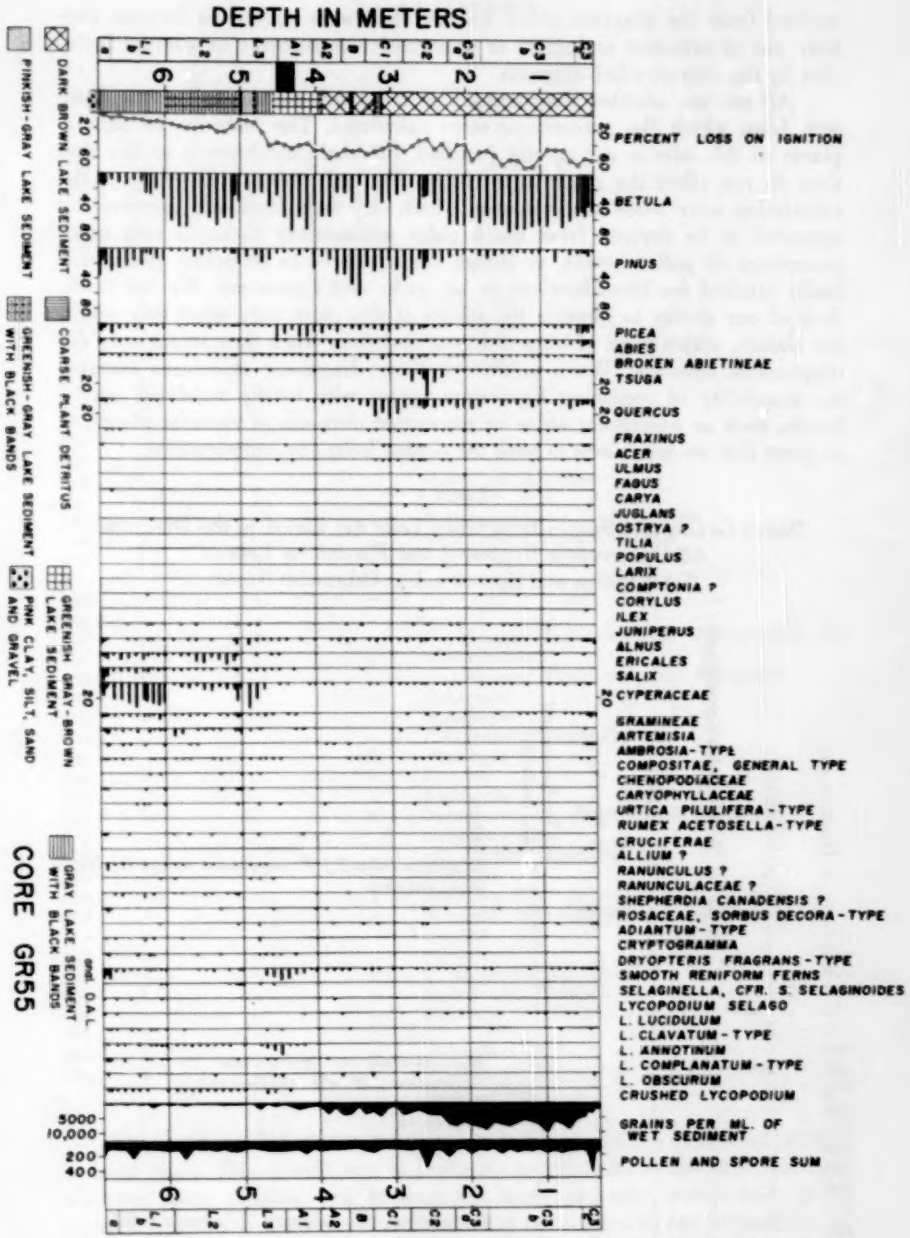

Fig. 1. Pollen diagram for Core GR 55, Gillis Lake, Richmond County, Nova Scotia. Black bar near left-hand margin indicates level of material for radiocarbon date Y-524—10,340 ± 220 years.

omitted from the diagram either because they were aquatic or because they were not of sufficient ecological or quantitative importance to warrant inclusion in the overcrowded diagram.

All species, whether diagrammed or tabulated, contributed to the basic sum from which the percentages were calculated. The inclusion of aquatic plants in this sum is not strictly justified, but their incidence is so low that they do not affect the result appreciably. The only things omitted from the calculation were some exotic spores, which by their form and preservation appeared to be derived from much older sedimentary rocks; a very small percentage of pollen grains, or things that appeared to be pollen grains, too badly crushed for identifications to be made, and *Equisetum*. We feel confident of our ability to identify the spores of this plant only when they retain the elators, which none of ours did, and uncertain when they retain only the diaphanous envelope. When neither of these diagnostic structures remains the possibility of confusing *Equisetum* spores with totally unrelated microfossils, such as unicellular algae or macerated elements of vascular plants, is so great that we felt it best to omit the counts from our computations.

TABLE 1

Pollen Grains and Spores from Gillis Lake not Listed in the Diagram
All Occurrences Represent one Percent or Less of
Total Pollen and Spores unless Otherwise Noted

Depth in cms.	% Isoetes	% Eriocaulon septangulare	% Nuphar	% Nymphaea	% Typhaceae and Sparganiaceae	Others
0	2	–	2	3	–	*Rumex acetosa-T.; Proserpinaca palustris; Hippuris vulgaris*
5	1	1	3	1	–	—
10	1	1	3	1	–	—
15	2	–	2	2	–	—
20	2	1	3	1	–	—
30	1	–	–	1	–	—
40	4	–	–	–	–	—
50	2	–	1	–	–	—
60	6	–	–	–	–	—
100	1	–	–	–	–	*Myriophyllum* cfr. exalbescens
110	–	1	–	1	–	Najadaceae; *M.* cfr. exalbescens
120	–	–	1	1	–	Boraginaceae
130	–	–	–	1	–	Najadaceae
140	–	–	–	–	–	Najadaceae
150	–	–	–	–	–	—
160	2	–	–	1	1	—
170	3	1	1	1	–	—
180	–	–	1	–	1	*Polypodium* cfr. *vulgare*; *"Claytonia"*—8%
190	2	1	–	1	1	—
200	1	1	–	1	–	Spergularia marina (?)

TABLE 1 (Continued)

Depth in cms.	% Isoetes	% Eriocaulon septangulare	% Nuphar	% Nymphaea	% Typhaceae and Sparganiaceae	Others
210	1	–	–	–	–	—
220	2	–	–	1	–	*Sphagnum*
230	2	–	–	1	1	"*Claytonia*"
240	–	1	–	2	1	*Lobelia*
250	2	–	–	–	–	Rosaceae
255	4	1	–	1	1	—
260	2	–	–	1	–	—
270	1	–	–	–	–	—
280	1	–	–	1	–	—
300	–	–	1	1	–	—
320	–	1	–	–	–	*Rubus* cfr. *hispidus*
330	–	–	1	1	–	—
340	–	1	–	1	–	Najadaceae; *Urtica* cfr. *procera*
350	2	–	–	1	–	Najadaceae; *Typha latifolia*
360	1	–	–	1	–	—
370	–	–	–	3	–	—
380	–	–	–	2	–	—
400	–	–	–	1	–	—
420	–	–	1	1	–	Najadaceae; *Typha latifolia*; Myriophyllum cfr. *alterniflorum*
440	–	–	–	1	–	
450	1	–	–	–	–	Najadaceae; *Typha latifolia*; *Sphagnum*
480	–	–	–	–	–	*Sphagnum*
490	–	–	–	–	–	*Sphagnum*
500	–	–	–	–	–	*Polygonum Bistorta-Type*
505	–	–	–	–	1	*Aquilegia* (?)
510	–	–	–	–	–	*Drosera*
530	–	–	–	–	–	*Sphagnum*
540	–	–	–	–	1	*Sphagnum*; cfr. *Matricaria maritima*
570	–	–	–	–	–	*Achillea-Type*
578	–	–	–	–	–	cfr. *Cornus canadensis*
590	1	–	–	–	–	*Plantago*
610	–	–	–	–	1	Rosaceae (?)
620	–	–	–	–	–	*Saxifraga oppositifolia-Type*; *Polypodium* cfr. *vulgare*
630	–	–	–	–	–	*Sphagnum*
640	–	–	–	–	1	*P. Bistorta-Type*; cfr. *Androsace*
650	1	–	–	–	–	*Sphagnum*; *Plantago*
680	–	–	–	–	–	*Plantago*; *P. Bistorta-Type*; Rosaceae (?)
682	1	–	–	–	–	—
684	2	–	–	–	–	*Sphagnum*
685	1	–	–	–	–	*Sphagnum*; Najadaceae; *T. latifolia*; *Epilobium*; Umbelliferae

The number of grains counted at each level was rather small—about 125 in most cases. This is sufficient to make the proportions of the principal constituents fairly reliable, but at least ten times as many would have to be counted in order to lend moderate reliability to changes in the proportions of the minor constituents. At a few levels peculiar spectra were found and two

or three times the usual number of grains was counted in order to see if these peculiarities were due to the small basic sum. In no case did any appreciable change result from increasing the sum to this extent.

Figure 1 includes a curve showing the total number of pollen grains and spores per milliliter of wet sediment. This curve is not very reliable, for although the quantity of material picked out for analyses was measured to about plus or minus twenty per cent, the extraction of pollen was not strictly quantitative. Only enough of the final concentrate was used to provide slides of a suitable density for counting and the rest was discarded. As a result, at the levels where pollen was scarce, particularly in the lower zones, the curve provides an estimate of the absolute pollen frequency that is of the correct order of magnitude. At the levels where pollen was very abundant, the value given is too low by some large but unknown amount. It should be noted that this error is one that makes the changes in the absolute pollen frequency appear less marked than they really are.

The core may be divided into three broad sections on the basis of its pollen content. In the basal section there is a high percentage of Cyperaceae, Gramineae, *Salix* and other indicators of open vegetation. This, following the usual and not completely satisfactory convention, may be considered the late-glacial part of the core, that is, the part deposited under essentially periglacial conditions. The center section lacks high percentages of open-vegetation pollen types, and is characterized by an abundance of *Picea*, *Abies* and *Pinus*. The upper section has in addition a variety of temperate hardwoods and *Tsuga*.

The bottom section can be sub-divided into three zones, which have been designated L1, L2 and L3, in accordance with the usage of Deevey (1951). L1 is a zone with a small amount of *Betula* and a moderate amount of *Pinus*, *Picea* and *Abies* pollen. It contains a great variety of indicators of open vegetation, of which sedges are quantitatively the most important. A tentative subdivision into a and b parts has been made for this zone on rather slender evidence. The L1a sub-zone has lower sedge and higher fern, Salix, and Alnus percentages than the L1b sub-zone. In addition it possesses a small trace of *Quercus* and a single grain each of *Epilobium* and Umbelliferae.

The change to zone L2 is a very distinct one. *Betula* rises sharply, *Pinus* falls, with its associated boreal conifers, and the incidence of indicators of open ground, particularly sedges, falls to low levels.

In zone L3 *Betula* declines, Pinus increases slightly and the indicators of open vegetation, particularly Cyperaceae, increase to levels comparable with those of zone L1.

Zone A1 is marked by an increase in *Betula*, a dubious decrease in *Pinus*, the virtual disappearance of indicators of open vegetation other than pteridophytes, and a dramatic rise in the incidence of *Picea*. In zone A2 Betula begins to decrease, *Quercus* is uniformly present in small amount, and Abies is present in quantities that are significant for such a weakly-represented genus.

The B zone is narrow and its borders are not sharply defined. It may be distinguished from A2 principally by the dominance of *Pinus* and from C1 by the sudden increase of *Quercus* and the appearance in small quantities of a number of hardwood trees, particularly ash, maple and elm in that zone.

In zone C2 oak is less abundant and there is a pronounced hemlock maximum, while pine declines steadily and birch rises. The upper zone, C3, is characterized by high birch and low hemlock and pine. It can be subdivided into C3a, with very low incidence of spruce and fir, C3b, in which these trees reappear, and C3c, in which *Betula* declines a little, spruce and fir increase, grasses increase, and a variety of indicators of open ground are present in amounts that are remarkably high for postglacial spectra. This is particularly true of a number of weeds: composites of various sorts and two kinds of *Rumex* conformable with the introduced European species *R. Acetosa* and *R. Acetoesella*.

VEGETATIONAL INTERPRETATION

Zone L1.—The presence of such a high percentage of sedge pollen in a lake facies seems to us very strong evidence that the L1 vegetation was open in character, with a good deal of the countryside in sedge meadow. The vegetation that comes to mind immediately is cottongrass tussocks, but we have not been able to identify cottongrass pollen in the profile. We have found one grain at Gillis Lake that matches our slide of *Scirpus caespitosus* quite well, and a few that we assign with more confidence to *Rynchospora alba*, primarily on the basis of texture. The great majority, however, we have been forced to lump into a general category that contains the bulk of our sedge collection and includes at least 17 species from the genera *Eleocharis, Dulichium, Scirpus, Kobresia, Cyperus, Carex* and *Cladium*. All of these are more or less pear-shaped in side view and have weak pores. There is considerable variation in the definition of the outline, some, like all our species of *Eleocharis*, collapse readily, others, like all our species of *Carex*, hold their shape rather well. It will be noticed that this group, to which most of our fossil grains belong, does not contain *Eriophorum*. If, as our limited experience suggests, cottongrass grains are always ellipsoidal with strongly protuberant pores, then the Gillis Lake sedge meadow was certainly not dominated by *Eriophorum* tussocks. We have only two reference slides of cottongrass but they are of *E. angustifolium* and *E. spissum*, two of the principal arctic tussock-building species at the present time.

Further support for the hypothesis of open vegetation is afforded by the high incidence of ericaceous shrubs in zone L1. The curve includes all of the *Ericaceae* (sens. lat.) that have grains in tetrads, as well as all the *Empetraceae*. It would be useful to distinguish the genera involved, but for present purposes hardly necessary, as most are indicators of open vegetation.

Other indicators of open vegetation are the grasses, composites, chenopods, chickweeds, mustards and pteridophytes. If we further inquire about what sort of open vegetation was involved, the abundance of *Salix* and *Artemisia* is of particular interest. *Artemisia* may be of importance in the pollen rain of either tundra or steppe, but high incidence of *Salix* in the pollen rain is characteristic of tundra. Single grains of *Epilobium* and *Saxifraga oppositifolia*-Type support the idea of tundra vegetation. We have checked the *Epilobium* grain against all the native genera of Onagraceae and are reasonably certain of its identity. Until we become more familiar with the tricolpate-

striate grains of the native flora, however, particularly some of the maples, we will have reservations about the identity of the arctic saxifrage.

These indicators of tundra conditions are all ones that might be found in low-arctic vegetation, although the incidence of *Artemisia* and *Salix* is such as to suggest middle- or high-arctic conditions. On the other hand, there are many plants that one would hardly expect to find in the middle-or high-arctic, ones that are, indeed, more characteristic of sub-arctic vegetation. We have in mind particularly *Shepherdia canadensis*, *Alnus* and all the pteridophytes except *Lycopodium Selago*. In general the open vegetation seems to have been sub-arctic or low-arctic in nature.

The evidence of the tree genera is not so easily dealt with. We shrink from separating the birch species on the basis of pollen alone. This may be possible where one has only a small number of different species and there is a considerable difference between the pollen sizes of the species. In the present flora of Nova Scotia, however, there are six native species of birches. In late-glacial vegetation one might expect to find three more, and in thermal maximum vegetation perhaps one or two additional ones, for a total of eleven species of *Betula* alone. When one adds *Ostrya*, which has very similar pollen, and the host of other triporate betuloid grains that can be distinguished from *Betula* with more or less certainty, the problem of species identification becomes well-nigh impossible. To separate North American *Betula* pollen into species on the basis of size frequency measurements of only fifty grains from each of three species, as was done in a recent study of central Canadian vegetation, may mislead the uncritical reader and obscure a very serious identification problem. We do not see how it can add to understanding of past vegetation.

Without specific identification of the birch species it is at least possible that they are shrubs of a sort to be expected in low-arctic vegetation. The Abietinae, however, cannot be dealt with in this way. *Pinus* rises to over 30 percent in L1, and there are considerable quantities of *Picea* and *Abies* as well.

Taken at face value this might be accepted as evidence of boreal forest during L1 time. We are inclined for a number of reasons to doubt that it is a valid indicator of closed vegetation. The matrix here is very inorganic, of just such a nature as the deposits that have long been known to be contaminated with redeposited forest tree pollen in Europe. Any high incidence of forest tree species, particularly if it is in conflict with other evidence, is suspect for this reason. A glance at the absolute pollen frequency will show that the total quantity of pollen present per milliliter of sediment in Zone L1 is very small, so a rather small amount of redeposited pollen might be expected to loom large in the pollen percentages. If such redeposition has occurred, we should find warm temperate trees of the sort that might mave inhabited Nova Scotia during interglacial times, as well as pine, spruce and fir. There is a trace of such pollen, particularly of oak and ash, but occasional grains of other hardwoods as well. Our sample of basal drift slipped out of the sampler in the field, as mentioned above, so we cannot apply Iversen's (1936) quantitative correction for rebedded pollen. If the contaminating

material, however, has a pollen spectrum comparable to that of the C zones, and if the hardwoods are all secondary in L1, then we might reasonably subtract a certain amount of the coniferous pollen in L1 from consideration, but by no means all. The balance might be the result of long-distance transport if only one genus were involved, but this does not seem a completely satisfactory explanation when three are represented by significant numbers of grains. Later evidence, particularly from macrofossil remains, may cause us to change our opinion, but at the present time we are inclined to suspect that there were a few spruce, pine and fir trees living in Nova Scotia during L1 time. The low absolute pollen frequency and the high frequency of so many indicators of open ground are convincing evidence against a continuous forest cover, but they do not argue so strongly against the presence of scattered patches of trees in particularly favorable microenvironments. In short, our impression of the vegetation is that it was hemiarctic (Rousseau, 1952).

The sub-zone L1a may be only a statistical artifact and vegetational interpretation is perhaps best left until it has been found in a few other localities. The high willow frequency suggests an arctic tendency, and the minor *Quercus* peak suggests a high level of secondary deposition.

Zone L2.—The outstanding characteristics of this zone are the rise in birch and the decline in sedge and conifers. Willow declines, and the incidence of heaths rises somewhat. This suggests that shrubs had become more important in the vegetation, but the continued presence of a variety of open-vegetation pollen types shows that there was still a good deal of countryside not completely covered by birch and heath shrubs. The decrease in conifers may not reflect an important change in the status of this element in the vegetation. The conifer pollen, whether derived, wind-blown, or the product of local stands, was probably diluted by the higher local pollen production resulting from a change to shrub vegetation.

There appears to be a decline in the frequency of ferns and of other pteridophytes except *Lycopodium clavatum* and *L. obscurum*. It may be noted, too, that *Artemisia* maintains itself at high levels and that *Abies* is depressed more than the other conifers. These phenomena suggest a dryer climate during L2 time than previously. The increase of shrubs at the expense of sedges also supports this view, as well as providing good presumptive evidence for a warming of climate.

Zone L3.—This zone appears to be characterized by a return to vegetation of the type that was prevalent in L1, but there are some differences. Birch does not fall quite so low, nor do the conifers rise so high. Particularly striking is the virtual absence of heaths from this zone, and the low values for pteridophytes that prevail in the bottom half of it. There may be some justification for recognizing two sub-zones on the latter basis, but it is well to wait for the phenomenon to be recognized in other cores before erecting a sub-zone for elements of such notorious local variability as these.

Zone A1.—The absence of indicators of open vegetation, coupled with the high values for spruce and birch, indicate that the vegetation of this zone was a closed boreal forest. The low incidence of *Abies* suggests that the climate was rather dry, but this is not supported by the rather high frequency

of fern spores. It may be that the seasonal distribution of precipitation was such as to afford an early winter snow cover for the ferns, but not an adequate supply of summer moisture for fir.

Zone A2.—This zone is distinguished from the preceding by the increase in fir pollen and also by the presence of *Quercus* in amounts that indicate the presence of oak trees in the forest. This suggests that the climate was somewhat warmer than it had been during the preceding time.

Zone B.—The very high incidence of pine pollen in this zone, coupled with a high absolute pollen frequency, must indicate a forest in which pine trees were important, although the heavy pollen production of this genus may obscure the importance of some of the other elements in the vegetation. The exact significance of this peak pine pollen must remain uncertain so long as it is impossible to distinguish the species involved. The possibility of using size statistics is better here than it is with the birches, but still attended by the possibility of grave error. Our material was made up by treatment with KOH and acetolysis, which introduces further difficulties, so we have been unable to determine the specific identity of our pine pollen grains.

Zone C1.—With the beginning of the C zones deciduous forest became more important than it was previously. This is shown not only by the high incidence of oak, but also by the presence in small but significant amounts of a number of other hardwood pollen types, particularly ash, maple and elm. This must have been a time of rather warm climate, judging from the amount of oak pollen present, an amount considerably greater than that present in the surface spectra.

Zone C2.—The outstanding characteristic of this zone is the very high peak of hemlock. This suggests a change to conditions that were somewhat moister than those that prevailed during C1 time and warmer than those of the present day. Such a pronounced peak of hemlock must indicate a forest in which it was an important constituent for this is not a tree that is so strongly represented in the pollen record as pine and birch are.

Zone C3.—With the decline of hemlock from the high values of the previous zone there is a corresponding decline in the other coniferous species. Birch is the dominant tree, and here again one wishes for information about the species that are involved. Even if the birch is *B. lutea*, the most southern one likely to have been an important forest constituent, this vegetational change indicates cooling from the conditions of the preceding oak and hemlock periods. The extremely low values of spruce and fir are something of a puzzle. They do not persist long, however, and in sub-zone C3b the trees return, fir first, then spruce. At the very top of zone C3 an important change takes place in the vegetation. Birch declines, while spruce increases. The absolute pollen frequency falls off sharply, and at the same time grasses become more important than they have been since late-glacial time. Other indicators of open ground also appear, either for the first time or in a frequency reminiscent of the L zones. Various composites are present in traces, and two species of *Rumex* are found in small amount. These last give an indication of the sort of open vegetation that is involved. All of the plants showing an increase in sub-zone C3c are weeds (under Cape Breton conditions *Picea* is the

most pernicious weed of the lot) but these two species compare very well with *Rumex Acetosa* and *R. Acetosella*, weeds that were first introduced into North America by white settlers. We do not have a complete reference set of native dock and sorrel pollen, but as there are no native species belonging to the same subgenera as these introduced ones we feel fairly confident that they are the sorrels present in C3c. We see no reason to doubt, even without this extra bit of evidence, that the open vegetation at the top of the core is due to the forest-clearing activities of European man.

STRATIGRAPHIC CORRELATIONS

In designating our pollen zones we have employed the system used by Deevey (1939; 1943; 1951) in New England and it may be well to draw attention to the ways in which we feel that our zones correspond to his, and the way in which they fit into late-Wisconsin stratigraphy generally.

The L zones seem to us to be temporally and vegetationally equivalent to those in Aroostook County, Maine (Deevey, 1951). So far only one climatic oscillation of the magnitude indicated by the vegetational shifts in our L zones is known from any place in temperate North America. There is very little question that the same climatic event is involved in the late-glacial oscillation at Gillis Lake, the one in northern Maine, and the one in the George Reserve in Michigan (Andersen, 1954). We believe that this event occurred at the same time in all three places, and that its vegetational expression was about as uniform as could be expected over so wide an area. Further, we see no reason to doubt that this event was the same as the one that lead to the classic stratigraphic sequence at Two Creeks, Wisconsin.

This correlation is supported by a radiocarbon determination by the Yale laboratory of gyttja from zone A-1 immediately above the L-3 boundary. This sample, Y-524, has a carbon age of 10,340 $\pm$ 220 years. Its position as indicated on the diagram is corrected for shrinkage of the core between the times of collection and carbon dating. The age is about 500 years younger than the age of the Mankato maximum, 10,856 $\pm$ 410 years (Flint, 1955). A second determination, of sediment from zone L-2, gave an age of about 13,000 years, or a thousand years older than one would expect for a Two Creeks deposit, but this date was not regarded by the laboratory as reliable, and the sample is being repurified for rechecking.

We feel that our A zones are comparable with the A zones of New England taken generally but the equivalence does not extend to the individual zones, not is it so well established as the L zone equivalence. For example, we are reasonably certain that A time in Nova Scotia began when it began in Maine, but we are not at all sure that it ended at the same time in the two places. We are quite certain that the A zones of southern New England, as originally designated (Deevey, 1939), are not contemporaneous with ours. Two carbon dates from southern New England (Leopold, 1956) agree in suggesting an age of about 13,000 years for the lower part of the A zone there.

The B zone at Gillis Lake appears to be stratigraphically equivalent to the B zone in southern New England, but we are not prepared to say that it

is temporally so. There is a certain amount of evidence (Flint and Deevey, 1951) which suggests that the B zone occupied a successively later time from south to north in eastern North America. This implies that B time in Nova Scotia was intermediate between B time in northern and southern New England, but the question is better left in abeyance until the pollen zones of eastern North America are more widely dated by absolute methods. Far from being certain that B zones of Nova Scotia and New England correspond, we suspect that the B zone may not even be vegetationally equivalent in Maine and Connecticut. In southern New England there is a certain stratigraphic distance between the decline of spruce and fir and the appearance of the mixed hardwood trees. Traces of oak are present in the B zone, and even in the A zones of some Connecticut profiles, but it does not enter in strength until the rise of *Tsuga* and the appearance of other hardwoods at what is incontestably the beginning of C1 time. This gap between the decline of spruce and fir and the rise of hardwoods contains a good deal of pine pollen, usually though not always with a peak on the pine curve.

The sequence of changes in northern Maine is rather different. There is no gap between the decline of spruce and fir and the rise of hardwoods. In some diagrams oak appears in strength before the decline of spruce and fir and it is soon joined by beech, elm and maple. Hemlock, however, does not appear with the mixed hardwoods, at least not in strength as it does in Connecticut, and in the gap between the decline of the boreal forest and the rise of hemlock the Maine diagrams show a well-developed broad pine maximum. It is this pine maximum, lying between the spruce decline and the rise of hemlock, that has been designated the B zone in Maine. We feel that there is some reason for regarding this as part of the C zones, for it is characterized by a mixture of hardwood species of weak pollen production which, though they are not impressive in the diagram, must have been very important in the forest. We have drawn our B and C boundaries at Gillis Lake in the light of this opinion. We feel that there is at least some vegetational equivalence between our B zone and the B zone in Connecticut, but perhaps not between our B zone and the B zone in Maine. As for temporal correlations, we should be rather surprised if the Maine and Nova Scotia B zones were contemporaneous, and not in the least surprised if the B zone at Gillis Lake were about as old as the B zone at Linsley Pond.

We recognize that this does not help the vexing problem of the pine zone, which was hard enough to understand before, and must be a good deal more so if it is to shift its position from B to C1. The essential difficulty is that the pine peak in Connecticut falls between spruce-fir and hardwood peaks, while the pine peak in Maine falls within the hardwood zone. The ecology of this situation must remain puzzling regardless of what arbitrary zone designations we apply, and regardless of when the events occurred.

Using our interpretation of the beginning of C1 it is very difficult to find a B zone in most of the Maine diagrams. We are not at all happy with this state of affairs and perhaps would be even less so if it were not possible to distinguish, in the Gillis Lake profile, one peak on the broad pine curve that does fall before the entrance of the hardwood species.

The Gillis Lake C zones seem to us stratigraphically equivalent to those of southern New England, but only as a group, not zone for zone. It appears that C3 shows some terminocratic tendencies, which would put the thermal maximum somewhere in C1 or C2 time, as it is in New England, but this is not a marked enough stratigraphic marker to be useful in precise temporal correlations. It may even be possible, considering the available body of evidence, that the thermal maximum was not contemporaneous throughout the world, although it is more likely that, like the accurately dated and definite Allerød oscillation, it was so.

In attempting a zone-for-zone correlation in this part of the column between Nova Scotia and New England one immediately runs into difficulties. To begin with, the *Tsuga* curve shows not a trace of the bimodality which is the basis of the C zone boundaries in New England. Instead of having two hemlock peaks, one in zone C1 and one in C3, we have a single hemlock peak in C2. Possibly one might infer that some local circumstance prevented the development of the second hemlock peak in Nova Scotia, and place the B/C boundary at the beginning of the Gillis Lake hemlock rise in accordance with Deevey's Maine practice, but then we create the anomally of a very well-developed oak maximum in B time. This seems to us even more Procrustean than the procedure we have adopted.

A case might be made for lumping the oak and hemlock zones together as sub-zones of C1, and calling our C3a, C2. This would bring the spruce curve into accord with New England usage but it would remove no other difficulty. It has the very serious disadvantage of applying a zone designation to a minor stratigraphic phenomenon and sub-zone designations to two major ones.

It may explain the central position of the single mode on the hemlock curve if we consider the very oceanic character of the Gillis Lake locality. In such a place hemlock may not have been depressed by drought at any time but may have been limited to a short period of dominance by temperature, so that it entered the pollen rain in force only at the very zenith of the thermal maximum. This explanation, however, confronts us with a new difficulty, for we have then to explain the presence of a great deal of oak during the previous time, which would have been cooler according to this interpretation.

Enough has already been said about difficulties of correlation with New England to make it abundantly clear that our C zones cannot be equated stratigraphically, let alone temporally, with those of Deevey. We should like, however, to draw attention to the fact already mentioned that the return of terminocratic species, particularly spruce and fir, in the upper part of C3 suggests very strongly a stratigraphic correspondence between at least part of this zone and C3 of Deevey.

The presence of three late-glacial zones representing the Allerød oscillation in Nova Scotia is of some geological significance. On the glacial map of North America (Flint, et al., 1945) Nova Scotia is mapped as undifferentiated Wisconsin. It is now possible to say that part of the Province was deglaciated prior to the Two Creeks interval. It is not possible to say just how long before, but the internal differentiation of zone L1 does not suggest a climatic

oscillation, so the presumption is that the last ice to cover the Gillis Lake area was of Cary (*sens. lat.*) age. This does not, of course, rule out the possibility that other parts of the Province have been ice-free for a longer time, or the possibility that parts of it have been covered by Valders or later ice.

DISCUSSION

One explanation that has been advanced to explain the presence in eastern North America of a number of vascular plants otherwise found only in the Cordillera, or with a few isolated stations in between, is that there was around the edge of the continental glacier a band of unforested land along which cold-tolerant herbs and shrubs could migrate. As Andersen has recently emphasized (1954) there is a paucity of stratigraphic evidence for such a corridor of migration. He could cite only the pioneering work of Deevey in Maine, which first disclosed evidence of late-glacial vegetation in America, and his own excellent study of a late-glacial deposit in the George Reserve. A preliminary account of another relevant study has been published by Leopold (1956). This is the fourth piece of published evidence for an open-vegetation corridor along the ice-front, and the first from the maritime area in which the Cordilleran disjuncts are found. Each such bit of evidence makes it more likely that the Cordilleran species may have had a continuous range across America during glacial times, but it takes positive fossil records of the species outside their present ranges to demonstrate that they actually did so. We cannot yet provide such records for the classic Cordilleran species, as Andersen (1954) has done for *Eleagnus*.

It is of interest to note the intermittent presence of *Artemisia* from top to bottom of the core. No species of *Artemisia* living in Nova Scotia today is regarded as a native (Roland, 1945).[1] The *Artemisia* grains in the late-glacial zones and A1 are probably species of northern distribution at the present time, perhaps *A. borealis*. The few grains in zones C1 and C2, however, can hardly be of northern species. It is likely that some southern species, such as the coastal *A. caudata*, reached *Nova Scotia* during thermal maximum time. The single grain in zone C3c may belong to any one of the introduced weed species that live in the Province today. This is one more example of the great mobility of plants and the plasticity of vegetation which have become so evident from detailed studies of Pleistocene microfossils. It should be remembered that palynologists become familiar with the pollen flora of their immediate surroundings first, so that exotic grains come to their attention despite general ignorance of the pollen morphology of the species concerned. A good deal of migration and extinction may be expected to escape notice for this reason.

The historical evidence dealing with the nature of early post-settlement Cape Breton Island vegetation has been summarized by Bentley and Smith (1954). Written accounts of the forests in the vicinity of Gillis Lake are few, and seem to have been over-influenced by the spruce-fir forest of the coast.

[1] Mr. Wilfrid Schofield has drawn our attention to some unpublished records of *A. borealis* in northern Cape Breton Island that demand the inclusion of this species in the native flora.

Only Bouchette, of the authors they quote, describes a forest for the land between St. Peters and Grand River which is in accord with the pollen spectra of zones C3b and C3c. It is interesting that Bentley and Smith (1954, p. 13), seem in doubt about the status of elm in Nova Scotia. Roland (1945) gave no hint of doubt that *Ulmus americana* is a native tree, and Auer (1930) included it in his pollen curves for mixed oak forest without comment. We have found it to be continuously present in the Gillis Lake section from the beginning of C1 time, and it must certainly be included in the native vegetation. Killam (1951) reached this conclusion from similar evidence.

Although these results have primary botanical significance their implications are of some zoogeographic interest as well. In essence, we have demonstrated that the Maritime Provinces have had a longer ice-free history than the central part of the continent at the same latitude. The poverty of the fish-fauna, remarked on in a previous paper (Livingstone, 1951) cannot be attributed to the short span of post-glacial time, which has actually been longer here than in parts of the Great Lakes region with a very rich fish fauna. The lack of suitable migration routes, perhaps combined with marine isolation during times of higher sea level, must be responsible.

Question about the nature of the late-glacial mammal fauna arises when one contemplates the long period of open vegetation. Presumably caribou would have been at home in such an environment, but the quantity of grass pollen in the L zones at Gillis Lake, and even more markedly in Maine, suggests the possibility of other herbivores, musk-ox and bison certainly, and perhaps even horse and mammoth. Pleistocene mammal remains are seldom reported from Nova Scotia, partly because there has been little interest in what appeared to be a most unpromising region. The few elephant remains that have been recorded (Piers, 1914; Livingstone, 1951) are all of the boreal forest mastodon rather than the grassland mammoth, and these have been badly rolled specimens of presumed interglacial age from gravel deposits. Reconnaissance of peat and diatomite diggings and roadcuts through Pleistocene sedimentary deposits may be expected to turn up remains of a variety of mammals now absent from eastern Canada, if European experience is a reliable guide. (Degerbøl and Iversen, 1945; Mitchell and Parkes, 1949).

The combination of open vegetation, whatever the animals that ate it, and marine food resources suggests that Nova Scotia deserves careful exploration for late-glacial human cultures. So long as it appeared that deglaciation was late and immediately followed by boreal forestation there was no reason to expect any cultural development markedly different from that of the historic Indians. It is now evident that a rather large space-time niche existed in the Maritimes for the development of a very different kind of hunting and fishing culture and archaeologists might hope to uncover some traces of it. One must, of course, recognize the probability that the principal settlements of late-glacial time were located along the contemporary sea-coast, and that during late-glacial time eustatic sea-levels were low because of the amount of water still locked in continental ice sheets. If isostatic rebound of the Maritimes was rapid then the late-glacial strand lines may now be buried under many fathoms of water and nothing but temporary hunting campsites may be ex-

pected on present dry land. If isostatic rebound was slow the late-glacial strand lines may be accessible today. This is a matter we hope to settle in future studies of lakes near present sea-level.

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ORTHOCLASE AND MICROCLINE GRANITES

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ABSTRACT. Some granites contain orthoclase, others only microcline. The former are mainly postkinematic, and the latter are mainly synkinematic and late kinematic granites. Orthoclase may form when emplacement is relatively rapid, microcline when it is more protracted.

INTRODUCTION

In earlier papers (Marmo, 1956a, b), the author has stressed the need to characterize granites before the problems connected with them can be satisfactorily discussed. The purpose of this paper is to discuss the fact that some granites contain predominantly orthoclase whereas others contain exclusively microcline as their potash feldspar. In particular, the presence of orthoclase is especially characteristic of the younger granites of alpine mountain chains and of rapakivi granites. The microcline granites are typical of Precambrian areas, in which orthoclase granites are more or less sporadic.

ORTHOCLASE OR MICROCLINE

Orthoclase and microcline are two modifications of potash feldspar. The former is monoclinic, and it is also formed if potash feldspar is crystallized in laboratories, but triclinic microcline has never been produced by direct crystallization. Indirect synthesis of microcline, however, has been reported. Wyart and Sabatier (1956) produced microcline by replacement of Na by K in an albite under hydrothermal conditions. In their successful experiment, the formation of microcline was expected, because the framework of the replaced albite was triclinic, and in this framework Na was replaced by K. So far, there is no satisfactory explanation why in direct crystallization the monoclinic form of potash feldspar is formed and not the triclinic, although the triclinic corresponds to the lower free energy level.

Many authors consider that microcline is an "alteration product" of orthoclase—that orthoclase, if heated for a long time at a temperature suitable for the rearrangement of the Al-Si framework of potash feldspar from monoclinic to triclinic, will re-arrange itself into microcline, and that this rearrangement can be entirely complete. Laves (1955), on crystallographic grounds stated that at least the microcline showing the typical cross-hatching must originally have been monoclinic, and consequently must have crystallized as orthoclase. On petrologic grounds, Marmo (1955) claimed a triclinic ancestry for the microcline of the Precambrian microcline granites (which contain no traces of orthoclase). He admitted, however, that in granites still containing orthoclase the microcline has probably originated by the rearrangement of the Al-Si framework of primarily formed orthoclase. To avoid controversy with Laves, Marmo (1955c) suggested that the symmetry of the molecular cell of the growing potash feldspar may initially have been monoclinic, but that it may have become triclinic at a very early stage of the growth of potash feldspar crystal, leading to the crystallization of triclinic microcline. This seems quite possible if suitable conditions and a long enough

time are available. Later, in a private communication, Laves admitted this possibility to the author.

In regard to the granite problem, it is very important to know which modification of potash feldspar originally grew in granites. If there are granites in which the primary potash feldspar was orthoclase and others in which it was microcline, then these granites have certainly had somewhat different origins or, at least, conditions of emplacement. If, on the other hand, it can be proved that in all cases potash feldspar has grown as orthoclase, all microcline granites must be considered as metamorphic granites. In the opinion of the author, however, both kinds exist: those with primary orthoclase, and those with primary triclinic microcline.

ORTHOCLASE GRANITES

In the Precambrian granites of Finland, as Mäkinen (1917) has pointed out, orthoclase does not occur but the potash feldspar is nearly always microcline. A few exceptions are known, however; two Precambrian granite groups containing much or predominant orthoclase: the granulite granites of Finnish Lapland (Eskola, 1952) and the rapakivi granites. In both, the potash feldspar is mixed and inhomogeneous. It is mainly orthoclase with an axial angle varying between wide limits; even in different parts of the same crystal 2V may have values between 50° and 75°. In such orthoclase, patches of well-twinned microcline commonly occur, and this microcline is undoubtedly a product of microclinization of orthoclase, as Eskola (1952) explained for the granulite granites, and Neuvonen (oral communication) for the rapakivi granites. Typically, the orthoclase in both types of granites is intensely perthitic (mainly hair perthite and cryptoperthite).

It is important to note that the young granites of the alpine mountain chains, normally pink and obviously postkinematic, are mainly orthoclase granites. The author, together with F. Permingeat of the Geological Survey of Morocco, has studied the pink, aplitic granite of Azegour from the Atlas. This is an orthoclase granite, and its potash feldspar is very inhomogeneous optically; it contains patches of microcline, much crypto- and hair-perthite, and in all respects it is very similar to the potash feldspar of rapakivi granite of Finland. Similar pink granites occur in other parts of the Atlas as well.

Orthoclase granites similar in all respects have been described from the Alps (Gysin, 1948), and from the Himalayas in Nepal (Gysin, 1956). The inhomogenetity of orthoclase and its joint occurrence with microcline has also been observed in the syenite (accompanying the iron ore of Gora Wysokaja) and the rapakivi of the Urals. Tuttle and Keith (1954) have described similar granite of Tertiary age from England.

All the granites mentioned above are typically intrusive. With the exception perhaps of the granulite granite, these rocks have mainly been regarded as postkinematic, emplaced after the main orogenic activity had ceased. In chemical composition, the orthoclase granites are rich in potassium and close to the ideal granite.

MICROCLINE GRANITES

In outer appearance the Precambrian late kinematic granites are commonly very similar to the younger orthoclase granites described above. Mostly they behave in the field as true intrusive bodies, forming dikes and sills with sharp contacts. Neither in hand specimen nor in the field can they be distinguished from the alpine, postkinematic orthoclase granites. Under the microscope, however, they differ from them very clearly: the Precambrian late kinematic granites contain no trace of orthoclase, their only potash feldspar being microcline, generally almost perfectly triclinic; also perthite, typical of the orthoclase granites, is much less common in the late kinematic microcline granites. In such granites in Sierra Leone, Finland, Sweden, and Russian Karelia, examined by the present author, perthite commonly occurs sporadically.

As mentioned at the beginning of this paper, some petrologists consider the microcline of these granites the product of microclinization of orthoclase. There are strong arguments against this opinion. Both in the alpine mountain chains and in the Precambrian granulite and rapakivi areas, dikelets of microcline aplite cut and hence are younger than the orthoclase granite. The aplites contain perfectly triclinic microcline but no traces of orthoclase, and the contacts of all these dikelets against the orthoclase granite are perfectly sharp. If the microcline of the aplites is the product of re-arrangement of the orthoclase, it is very difficult to explain why this microclinization proceeded to a small extent only in the orthoclase granite but to completion in the younger aplite enclosed by the orthoclase granite. If, as argued above, microcline can be primary, the facts are easier to explain by supposing that the potash feldspar of the aplites is such primary microcline.

Thus the author is inclined to classify the younger granites into microcline granites and orthoclase granites. The former contain only microcline of high triclinicity; the latter contain orthoclase, partly re-arranged (orthoclase of varying order in its Al-Si framework), to some extent also microclinized, but still always present.

DISCUSSION

Actual intermediates between orthoclase granites and microcline granites (as defined above) have not been seen by the writer. So the orthoclase in orthoclase granites is inhomogeneous, the relative amount of microcline may, of course, vary largely, but it was mainly formed from the orthoclase, which is invariably present. Furthermore, minute veinlets of microscopic size composed of highly triclinic microcline may occur commonly in the orthoclase granite, but these veinlets are always definitely younger than the main part of the rock. Apart from this microcline, the varying triclinity of the potash feldspar of the orthoclase granites is due to the microclinization of monoclinic potash feldspar. Microcline granites, on the contrary, contain microcline of high triclinity but no orthoclase.

Petrochemically the two granite types are very similar, and both approach the composition of ideal granites. Likewise in the field they are very similar; they are intrusive and form similar bodies, mainly veins and dikes,

and they tend to be aplitic, commonly pink, fine- to medium-grained rocks. Obviously, both granites were derived from similar material but were emplaced in different ways or under different conditions.

As mentioned above, the orthoclase granites are mainly postkinematic granites, many of them Paleozoic intrusives. In Precambrian areas they are much less common, and are actually restricted to granulite granites (in Finland) and rapakivi granites (Finland, Ukraine, the Ural), which also have been considered as postkinematic. The microcline granites are especially typical of Precambrian areas and are mainly late kinematic. Furthermore in synkinematic granites in which the potash feldspar was mostly introduced metasomatically (e.g., Eskola, 1956) it is exclusively microcline. This applies to both Precambrian and Paleozoic synkinematic granites, as far as known to the author. Reports in earlier papers of orthoclase in Finnish synkinematic rocks have been disproved by more recent new investigations. Thus orogenic conditions (both syn- and late kinematic) seem to be particularly suitable for the growth of microcline, whereas postkinematic conditions favor orthoclase. But even then primary microcline can crystallize, as in the aplite dikes that cut the orthoclase granites.

According to the author (Marmo, 1955b), if potash feldspar crystallizes sufficiently slowly and at a temperature lower than that at which the triclinic form is replaced by the monoclinic form, microcline will grow instead of orthoclase. If, however, either the temperature is somewhat elevated, or the duration of crystallization is short, orthoclase will grow instead of microcline. Probably, mainly depending on the duration of crystallization, the triclinicity of potash feldspar will vary, rapid crystallization (and rapid accumulation of crystallizing material ?) favoring most nearly monoclinic orthoclase. Consequently, synkinematic and late kinematic crystallization of potash feldspar will yield microcline because it could take place in an appropriate temperature and because it is there sufficiently sluggish.

Late kinematic orogenic movements must have been very sluggish, and there is no reason to believe that deep-seated intrusions of granite would have taken place promptly. Furthermore, late kinematic granite intrusions probably took place under hydrothermal conditions and not as dry magma. Thus in the granites emplaced at this stage, microcline would be formed and not orthoclase. Postkinematic granite intrusions, on the other hand, probably followed already open fractures. Therefore intrusion at that time could be faster, and orthoclase could develop instead of microcline.

After the main postkinematic orthoclase granite intrusions, some material of similar composition probably still remained, not yet incorporated into the orthoclase granites. This material collected in fractures in the already solidified granite, not so much by intrusion as by slow infilling of the fractures by granitic material. In these fractures, slowly occupied by aplite, primary microcline could grow instead of orthoclase, producing the microcline aplite dike cutting the orthoclase granite.

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DISCUSSION

FURTHER COMMENTS ON STRATAL TERMS

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Professor Schindewolf's recent paper (1957) calls for a few further remarks. I am glad that, in the process of "correcting my apparent misconceptions" (*ibid*, p. 394), he has been led to concede the desirability of abandoning the old German use of the word Formation in favor of System, and of regarding a Zone as a band of rock (*ibid*, p. 356, 357). I am delighted to see him writing of correlations on several pages without inhibition. As regards Zones, despite his concession, Professor Schindewolf does not sound convinced. Even if my quotations from Oppel's works were "lengthy" and my interpretation of them was "apodictic", I should have thought that no one could deny that they proved in Oppel's own words that he considered that zones, "could overlie or underlie other zones, that they could be the lower or upper half of a Formation (*sensu anglico*), e.g. Ornatum Clays, that they could contain fossils, be exposed, occur in different facies in different places, and possess a limited area of distribution" (Arkell, 1956, p. 460). If that is admitted, I cannot understand how any one could argue that an Oppelian zone is a time-unit, though I never denied, and would not deny, that a zone "has a chronologic sense" (Schindewolf, 1957, p. 397).

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REVIEWS

Géologie sédimentarie. Les séries marines; by AUGUSTIN LOMBARD. P. 722; 180 figs., 13 pls. Paris, 1956 (Masson & Cie, 11,000 fr.).—Lombard's ambitious treatise is divided into four parts: I. Modern deposits of seas (and lakes); II. Marine deposits in the geological record; III. Principles of stratigraphy; and IV. Origin of marine deposits. Part I is a review of the factors that influence modern marine sediments. Part II presents some of the kinds of marine sediments, their stratification, criteria for superposition, and methods of analysis, emphasizing the lithologic curve and graphic presentation of the composition and structures. Part III. (*Les ensembles sédimentaires*) consists of a discussion of the principles of stratigraphy (correlation, facies, etc.) and the rhythmic arrangement of sedimentary deposits. Part IV considers the larger associations of sedimentary rocks as controlled by tectonic movements, with illustrations from the Alps, Appalachians, and Rocky Mountains. Sequences that have been analyzed by their lithologic curves[1] are emphasized throughout the book and inferences drawn about transgressions, regressions, and stability from the shape of the curve.

[1] The "lithologic curve" and plots of other parameters of sedimentary deposits are devices much used by some European stratigraphers. As such curves are seldom employed in the United States, the following brief summary is offered by way of explanation. The "lithologic curve" is constructed opposite a graphic columnar section drawn to scale. Parallel

vertical columns are laid off alongside the section, a separate column being reserved for each different lithotope present in the section. Coarse-grained fragmental rocks are assigned to columns on the left; lutites, in the center; and carbonate rocks, on the right. A point is marked in the proper column opposite the lithologic subunit concerned and the points connected by a straight line. The curve is not continued across known stratigraphic discontinuities. Study of the shapes of the curves provides a rapid means of visualizing vertical stratigraphic variations at a particular locality.

Though the reviewer recognizes the logic of Lombard's organization of the material, he feels this arrangement impedes coherent presentation of such topics as stratification and cross-stratification, "graded bedding" (*granoclassement* of Lombard), other sedimentary structures, and organic remains, to mention but a few. It is necessary for Lombard to consider these matters separately in several parts of the book, and he does so without cross reference to similar sections on other pages, and with considerable repetition of material and loss of continuity.

Lombard excels in his *stratonomie* (Part II); he presents here the most comprehensive discussion of strata and surface markings of stratal surfaces known to the reviewer and is particularly exhaustive in his treatment of rhythmic sequences, a subject of which Lombard is particularly fond. The symbols used to portray stratigraphic field observations graphically are models which deserve emulation (though they differ from a recent American attempt to achieve the same end). The subjects of disturbances of stratification and markings of stratal surfaces caused by current action, however, suffer by comparison; their presentation is confused, inconsistent, and in the reviewer's judgment, inadequate.

The author seems particularly sensitive on the role turbidity currents play in the distribution of sediment and in the origin of strata. He summarizes the experiments of Kuenen, the extension of these ideas to recent deep-sea research by many workers, and their application to the geologic record. Where these ideas have been applied to the geologic record Lombard is content to note in passing without comment that such application has been made; it does not seem to have altered his point of view. Throughout, one feels he hesitates; he regards turbidity currents as real, but without general significance. Lombard characterizes turbidity currents as rapid, local "accidents" of sedimentation and maintains that his concept of slow flow (*écoule lentement*) of sediment along the bottom differs in scale, speed, and volume from turbidity currents as the behavior of snow moving gently downslope under the sun differs from the rapid descent of an avalanche. Having rejected the turbidity-current mechanism, Lombard analyses strata in terms of the "geologic weight" (*poids géologique*) of the material composing them. Two categories are recognized: most terrigenous sediment larger than clay-size, which possesses such "geologic weight" (termed *pondérables*) and fine-grained clastic particles and precipitates from solution, which lack "geologic weight" (termed *impondérables*). The *pondérables* form strata by spreading themselves out along the bottom, guided in their progress by bottom topography; the *impondérables* blanket the bottom like falling snow. The reviewer fails to discover wherein the action of the *pondérables* differs from turbidity currents of low density and slow velocity.

The reader is reminded at the very beginning of the limitations of narrow adherence to the Lyell-brand of Uniformitarianism, as has been argued previously by Cayeux. This may seem like excess baggage to most American readers, who generally view their Uniformitarianism in the broader light of the constancy of physical and chemical laws, rather than in a restricted geographic context.

Lombard's new classification of areas of marine sedimentation consists of: 1) continental platforms (subsiding or stable); 2) intercontinental-, marginal-, of intracratonic marine basins (subsiding or stable); 3) trenches and cordilleras; and 4) oceanic plateaus and basins. He concludes all marine sedimentary deposits can be understood by superposing onto these larger entities the products of euxinitic, paralic, lagoonal, and/or tectonic environments. The reviewer regards this as a sound approach to the larger analysis of sediments, for it includes both the basin *and* the several depositional environments which may occur in more than one type of basin. Lombard's scheme, however, may fail for want of categories. In perusing the examples cited in Part II, Chapter 1, the reviewer was startled to find the non-marine Triassic Newark group of the eastern United States assigned to Group 3 (trenches and cordilleras).

American workers in sedimentary rocks will find Lombard's book a useful, though somewhat awkwardly organized and carelessly edited treatise on European, particularly Alpine, stratigraphy. The cosmopolitan bibliography covers 34 pages, but lacks reference to Garrells, Marcus Goldman, Mackie, Lyell, Rubey, or Sorby, names familiar to English-speaking students of sedimentary rocks. Many citations of author and date stand naked in the text without full listing in the bibliography. The literature on rhythmic deposits is treated extensively.

JOHN E. SANDERS

The Living Rocks; by STEVAN CELEBONOVIC, with commentary by Geoffrey Grigson. P. 94. New York, 1957 (Philosophical Library, $6.00).—This collection consists of 64 full-page black and white photographs of outstanding specimens of minerals and fossils from the collections of the museums in Geneva, Paris, and London, which have been selected for beauty of form. Some of the specimens resemble objects of modern art, especially sculpture, to which they are compared. The text dwells upon the wonder as well as beauty of natural objects, both of which tend to be forgotten as they are analysed in more and more detail by the scientist.

The photographer, Célébonovic, introduces his Art and Nature series most auspiciously with this excellent volume. Its pages should be perused with interest and pleasure by any geologist. Some of the photographs should challenge the experts ability to recognize common minerals without the usual cues of color and luster. The book might well attract the interest of non-geologists to the fascinating materials of the earth sciences.

JOSEPH T. GREGORY

Détermination Pratique des Fossiles; by ANDRE CHAVAN and ANDRE CAILLEUX. P. 388; 586 figs. Paris, 1957 (Masson et Cie; 5,000 fr. or 5,800 fr. hard back).—This is the third in a series of keys to geological objects prepared by the authors for amateur naturalists. The other two were about rocks and minerals. The book is divided in two parts. The first 164 pages are devoted to simple morphologic keys to some 882 fossil genera; including 707 invertebrates, 109 vertebrates and 66 plants. In part two the genera are described, and related genera noted, under a much simplified systematic arrangement. About half the genera are illustrated by line drawings.

The book is intended for the western European amateur. It will be of interest to American amateurs who read French and have European specimens in their collection. As a general guide for North American fossils it is of limited use. Many of the common European genera treated are unknown or very rare in this country; conversely, many of our common fossils are not included because they are rare in Europe. KARL M. WAAGE

Acoustics; by JOSEPH L. HUNTER. P. xv, 407; 221 figs. New York, 1957 (Prentice-Hall, Inc., $8.50).—This book is written on a level for those with a background of general physics and elementary calculus. The first part of the book is an introduction to acoustic theory; the second part considers applications of the theory to several subjects of present research interest.

The first four chapters, which make up the introductory portion, contains most of the mathematics and important acoustical concepts needed for the latter portion of the book. The liberal use of figures and of illustrative examples make the text easy to follow. The writing is, however, rather uneven; several derivations, such as that for the wave equation for the string, are too short, while more elementary material is given in considerable detail. Moreover, there are several important omissions in these chapters, so that the reader does not get the "solid introduction to fundamental acoustics" desired by the author: for example, the derivation of the acoustical wave equation is done on a very elementary basis; also, no mention is made anywhere of the existence of shear waves.

The second part of the book consists of eight chapters which serve as an introduction to the fields of loudspeaker and microphone design, recording, acoustic measurements, speech, architectural acoustics, ultrasonics, and underwater sound. The chapters on loudspeaker and microphone design are well done; considerable material is given concerning design techniques and the compromises necessary to achieve satisfactory response. Reciprocity calibration of microphones is discussed in the chapter on acoustic measurements; also included is a short summary of the characteristics of several instruments used in measurement of loudness and frequency. The chapters on speech and architectural acoustics are probably the best in the book; the author not only gives an introduction to the important concepts, but explains why they are important and what led to their adoption. The chaper on ultrasonics (of fluids) covers Stokes' classical theory of sound absorption and the theory of molecular relaxation. Echo ranging, a qualitative description of refraction, and underwater transducers are described in the underwater sound chapter.

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